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THE MONROE DIVISION OF ROCKS IN OHIO[1]

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ABSTRACT

The Monroe division of rocks around the west end of Lake Erie has long been considered of Silurian age and was especially so interpreted by Grabau (Michigan, 1910). Later Stauffer (1916) and Williams (Ontario, 1919) assigned the Upper Monroe to the Devonian. The present paper, based upon a study of the Monroe of Ohio, presents the several lines of evidence bearing on the age of the Monroe and concludes that, on the bases of geographical distribution of the members, stratigraphical relations, and faunas, the Upper and Middle Monroe are of Devonian age, the systemic contact in Ohio being at the base of the Sylvania (Middle Monroe) sandstone.

INTRODUCTION

The Monroe division is at the horizon of the Silurian-Devonian contact, and this paper bears upon the problem of the location of this systemic contact. The Monroe unit is referred to in this paper by the indefinite term of "division" because as here interpreted it will not fit any of the more exactly defined subdivisions such as "formation" or "series."

The Monroe division underlies an area of several thousand square miles in western and northwestern Ohio, as shown by the combined shaded areas of Figure 1. It is also the bed rock for a part of Monroe County in the southeast corner of Michigan and for the southwest tip of the Ontario peninsula, and it underlies the west end of Lake

[1] Published with the permission of the director of the Geological Survey of Ohio.

Erie west of Pelee and Kelleys islands. It continues westward into Indiana, where it has a small areal distribution.

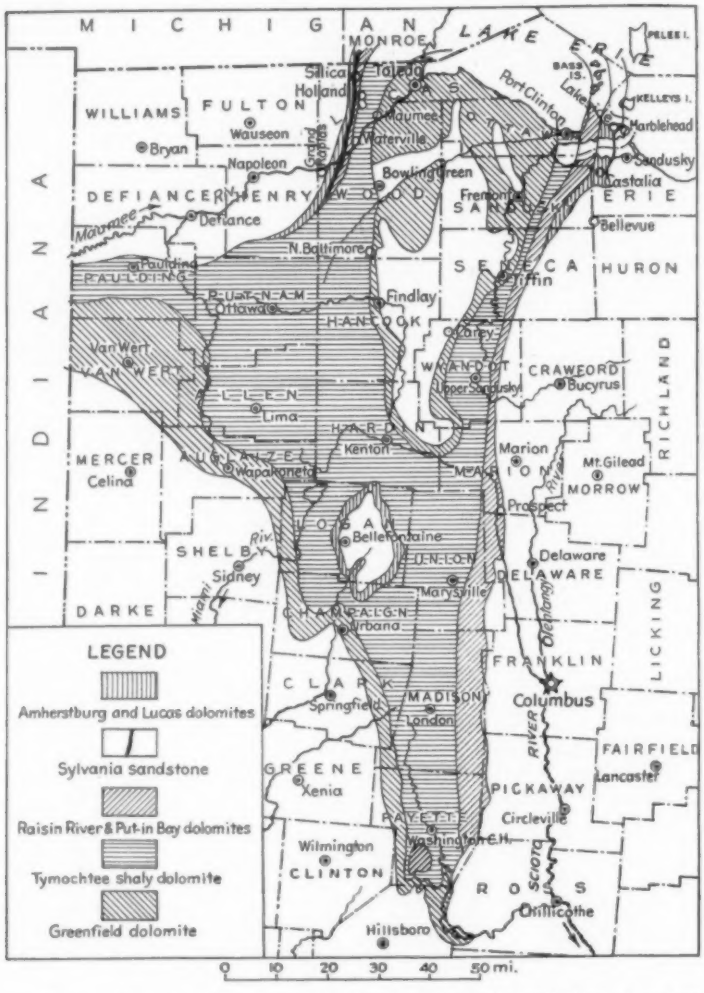

FIG. 1—Map of northwestern Ohio showing the areal distribution of various subdivisions of the Monroe.

The first names applied to the rocks in Ohio which we now know as Monroe were given by Newberry in 1869,[1] being taken from the geologic column of New York (see Table I, col. 1). These names included (1) Salina, for the gypsum-bearing strata on Sandusky Bay which he thought must be related to the salt beds of New York; (2) Waterlime, recognized by its fossils on South Bass (Put-in-Bay) Island; (3) Oriskany, for the sandstone in Lucas County; (4) Corniferous, for all the limestone above this sandstone.

These names were used in various publications of the Geological Survey of Ohio until 1888, when Dr. Edward Orton showed[2] that the gypsum-bearing strata of Ottawa County are about 400 feet above the horizon of the Salina, and he extended the Waterlime division down to include these strata. Further, he showed that the basal part of what had been called the Corniferous in Lucas County was not Corniferous, and this also he assigned to the Waterlime and thereby also placed the sandstone of Lucas County, which had been called the Oriskany, in the Waterlime. To this sandstone he gave the name Sylvania. The Waterlime, thus extended downward and upward by Orton, comprised a thickness of about 600 feet and included the Salina, Waterlime, Oriskany, and the lower part of the Corniferous as these terms were used by Newberry for Ohio (see Table I, col. 2). It included all the strata of western Ohio from the top of the Niagara to the base of the true Corniferous, which we now call the "Columbus limestone."

In 1895, Dr. A. C. Lane, state geologist of Michigan, introduced[3] the term "Monroe beds" for the rocks of southeastern Michigan "extending from the limestones of the overlying Dundee (Columbus) down to the lowest gypsiferous beds." This term, as introduced, included not only what had previously been called Waterlime in Michigan and Ohio, but also the underlying salt containing strata of southeastern Michigan, which were recognized as the equivalent of the Salina of New York. As defined, the term Monroe included everything from the top of the Niagara to the base of the Dundee (Columbus). The lower part, the Salina, was later removed from the Mon-

[1] *Geol. Surv. Ohio, Rept. of Progress in 1869*, pp. 15–18.

[2] *Geol. Surv. Ohio*, Vol. VI, pp. 15–20.

[3] *Geol. Surv. Mich.*, Vol. V, Part II, pp. 26–28.

roe.[1] In 1903, Dr. Prosser used the term Monroe[2] for those rocks of western Ohio lying between the Niagara and the Columbus, and further divided the Monroe into three members, the middle one of which is the Sylvania sandstone of Orton. For the strata below the Sylvania he used the term Tymochtee, which had been used by Winchell in 1873 for a part of the Waterlime, and for the strata above

TABLE I

CORRELATION OF THE NAMES USED BY VARIOUS AUTHORS FOR DIVISIONS OF THE ROCKS NOW KNOWN AS MONROE

Newberry 1871	Orton 1888	Prosser 1903	Lane, Prosser, Sherzer and Grabau, 1907	
CORNIFEROUS	CORNIFEROUS	COLUMBUS	COLUMBUS LIMESTONE	
		LUCAS	DETROIT RIVER (Upper Monroe)	LUCAS DOLOMITE
				AMHERSTBURG DOLOMITE
				ANDERDON LIMESTONE
				FLAT ROCK DOLOMITE
ORISKANY	WATERLIME / SYLVANIA	MONROE / SYLVANIA	MONROE / SYLVANIA SANDSTONE (Middle Monroe	
WATERLIME		TYMOCHTEE	BASS ISLAND (Lower Monroe)	RAISIN RIVER DOLOMITE
				PUT-IN-BAY DOLOMITE
SALINA				TYMOCHTEE SHALY DOLOMITE
				GREENFIELD DOLOMITE
NIAGARA	NIAGARA	NIAGARA	NIAGARA	

Sylvania and below the Columbus limestone he proposed the name "Lucas limestone" (see Table I, col. 3).

In 1907 there appeared a much more detailed classification of the Monroe, proposed in a short paper by Lane, Prosser, Sherzer, and Grabau.[3] This classification, which is given in the last column of Table I, was the outgrowth of a study by Grabau and Sherzer of the Monroe of southeastern Michigan,[4] the most comprehensive study of the Monroe that has been published. This classification of 1907 has been in general use since that time.

[1] A. W. Grabau, *Science* (N.S.), Vol. XXVII (1908), p. 622.

[2] *Jour. Geol.*, Vol. XI, pp. 538–41.

[3] *Bull. Geol. Soc. Amer.*, Vol. XIX, pp. 553–56.

[4] *Mich. Geol. and Biol. Surv.*, Pub. 2 (1910). Pp. 248, pls. 32.

The Waterlime and the Monroe have long been considered of Silurian age. In the work by Grabau and Sherzer on southeastern Michigan the Silurian age of all the Monroe was confirmed, though emphasis was placed upon the Devonian aspect of the faunas of the Anderdon and Amherstburg members of the Upper Monroe. In 1916, Stauffer reviewed the stratigraphical relations existing above and below the Detroit River series,[1] and showed the close relation faunally of the Amherstburg to the Onondaga of the Devonian, and in a report on the Silurian of the Ontario peninsula published in 1919, Williams[2] placed the Upper Monroe, or Detroit River, in the Devonian on a faunal basis. The conclusions of the present paper are, in general, in accord with the conclusions of Stauffer and of Williams. As interpreted by Williams and followed in this paper, part of the Monroe is Devonian and part Silurian, and therefore the term Monroe cannot be given any rank in a standard classification.

The region within which the Monroe exists is drift covered, and rock outcrops are few, scattered, and commonly small. It is therefore not possible to trace horizons or zones for any considerable distance. There are very few continuous vertical sections of as much as 50 feet, and only rarely can separate sections be exactly correlated so as to make continuous sections of greater thickness. The total thickness of the Monroe is about 600 feet, but a continuous section of this cannot be compiled. The various members have their greatest development at different places within the area, and the sum of the maximum thicknesses of the members would probably total over 1,000 feet.

The Monroe as a whole may be described as a drab, thin-bedded, laminated, compact, argillaceous dolomite. However, there is much variation, and although rock fitting this description may be found in all the members, each member also has one or more other types of stone within it. There are zones of massive porous dolomite, of evenly bedded non-laminated dolomite, of brecciated dolomite, of dolomitic shale, of limestone, of sandstone, and of gypsum.

The Monroe as a whole contains few fossils, and these are commonly of small size and consist of external and internal molds. The

[1] *Bull. Geol. Soc. Amer.*, Vol. XXVII, pp. 72–77.

[2] *Geol. Surv., Canada, Mem. 111*, pp. 18–22.

number of species in the several members ranges from about fifty for the Amherstburg to less than five for the Put-in-Bay and the Tymochtee. Only rarely does one find a stratum in which fossils can be said to be common.

BASS ISLAND FORMATION

The name Bass Island was proposed in the classification of 1907[1] for that part of the Monroe below the Sylvania sandstone. The name is taken from the group of islands of that name in the western part of Lake Erie north of Ottawa County, Ohio.

GREENFIELD DOLOMITE

The name Greenfield is from the town of Greenfield in northeastern Highland County, near which place this member is exposed and has been extensively quarried. The Greenfield is a drab, fine-grained dolomite with carbonaceous partings and commonly in beds of 2–6 inches, although at places with thicker beds or massive ledges. This massive phase is rough textured and vesicular, with corals and stromatoporoids, and is a kind of reef rock which at places can be seen to pass laterally into the even-grained, bedded type. The thickness of the Greenfield is apparently between 75 and 100 feet, but a complete section is not known at any place.

Many exposures showing the contact of the Greenfield with the Niagaran below have been found in northern Ohio. A few of these indicate a slight disconformity, but most of the exposures, covering parts of Wood, Ottawa, Sandusky, and Seneca counties, show apparent conformity with very characteristic transition changes and with very widespread uniformity of the beds just below and just above the contact. No outcrop is known showing the relation of the Greenfield to the Tymochtee above.

The Greenfield, as in fact most of the Monroe, is characterized by undulating bedding with local dips of 1–3° in various directions, forming low domes and shallow basins 40–100 feet across, which are apparently original structures. There may be also small folds of secondary origin. The undulations are so irregular in arrangement, and the direction and amount of dip changes at such short intervals,

[1] *Bull. Geol. Soc. Amer.*, Vol. XIX, p. 554.

that no dependence can be placed upon elevation of contacts or rock layers in correlating strata from exposure to exposure.

The distribution of the Greenfield is shown on Figure 1. The type locality is in northeastern Highland and southwestern Ross counties in the exposures along Paint Creek and its tributaries. Farther south in eastern Adams County there are a number of small areas, but in general the Monroe is overlapped here by the Ohio shale. In this region from Greenfield southward to the Ohio River, only the Greenfield member of the Monroe is present on the outcrop. There are a few exposures of the Monroe in southern Fayette County, but in general they are of strata above the Greenfield member. Farther northward in Fayette and in Madison counties the rock is deeply drift covered and few exposures exist. The Monroe-Niagaran boundary extends north-northwest across Fayette, southwestern Madison and eastern Clark counties, and the Greenfield member should underlie here the western part of the Monroe area.

The Greenfield dolomite has an extensive development in northern Ohio around the borders of the Niagaran area which here forms the axis of the Cincinnati anticline, the dominant structural feature in the geology of western Ohio. The Greenfield is known near Carey, Findlay, and Tiffin, all of which are on or near the Niagaran-Monroe boundary. It forms the north-south belt of Monroe through Sandusky and western Ottawa counties, being known at many exposures, a number of which show the contact with the Niagaran below. Exposures at Ballville just south of Fremont have long been known as showing the typical Greenfield fauna. It also underlies much of southeastern and northern Wood County with many exposures, and probably much of eastern Lucas County, being exposed in the valley of the Maumee River at the village of Maumee.

The Greenfield is known at a number of places west of the Cincinnati anticline in western Allen, southwestern Putnam, and Van Wert counties and is probably the bed rock of much more of these counties where deeply buried by drift. It no doubt underlies a belt along the southwest border of the Monroe area through Auglaize, Logan, and Champaign counties.

The fauna of the Greenfield consists of about fifteen species. The most common forms are *Hindella rostralis* and *Leperditia ohioensis*,

which locally are abundant. Other characteristic species are *Schuchertella hydraulica*, *Leperditia angulifera*, *Camarotoechia hydraulica*, and *Rhynchospira praeformosa*.

TYMOCHTEE SHALY DOLOMITE

In describing the section on Tymochtee Creek in Crawford Township, Wyandot County, Winchell gave the name "Tymochtee slate"[1] to 24 feet of strata comprising Nos. 7, 8, and 9 of his section. However, the name soon came to be used for the entire section on Tymochtee Creek and then to include the thin-bedded phase of the Waterlime everywhere throughout northwestern Ohio. The classification of 1907 retained the term, but with the statement that it was "a tentative division of the Lower Monroe" possibly not as distinct as the other members. The rocks of the Tymochtee Creek exposure are gray to drab, thin-bedded dolomite, with partings or thin layers of carbonaceous, argillaceous material. Similar strata are found at various other places over the Monroe area, indicating a division of rather wide extent. These various outcrops cannot be traced into each other, and none of them can be traced into the Tymochtee Creek exposure, but the general similarity of the stone is such that it should all be called Tymochtee until something definitely against such an interpretation is determined. At none of these exposures is the superposition of the Tymochtee above the Greenfield proved by outcrop, but the geographical position of the outcrops is in every case in accord with such an interpretation. The Tymochtee member is drab, thin-bedded, laminated, argillaceous dolomite with much carbonaceous material as partings along the bedding planes. It is the most shaly member of the Monroe and at places contains distinct zones of shale. The thickness is about 150 feet, but no complete exposure exists.

The Tymochtee member has its greatest development in the middle part of the Monroe area, being widely distributed over the axis of the anticline in western Hancock, Allen, Hardin, Logan, and Union counties (Fig. 1). Farther northward it forms a belt on either side of the axis, continuing to the lake border. It is present at lake level on the west of South Bass and Catawba islands, where it pre-

[1] *Geol. Surv. Ohio*, Vol. I, Part I (1874), p. 633.

sents a very shaly phase overlaid directly by the Put-in-Bay member. On the west side of the axis it is present in a belt through western Wood and Lucas counties, with its best exposures in quarries at North Baltimore, Wood County, and near Waterville and Holland, Lucas County. In these last two quarries there are strata far more shaly than the usual Tymochtee farther south, and in the Holland quarry these are directly overlaid by strata that represent the Put-in-Bay member.

The stratigraphical relation of the Tymochtee to the Greenfield below is somewhat uncertain. There are numerous exposures of strata that are typical of either member, but there are no exposures showing typical Tymochtee and typical Greenfield in the same section, and there are also a number of exposures that one hesitates to assign positively to either member. There are also various degrees of the shaly character within the Tymochtee. It has indeed been suggested that the Tymochtee is a facies of the Greenfield; that it passes laterally into the typical Greenfield and is therefore the time equivalent of a part or all of the Greenfield. In the classification of 1907 the Tymochtee was referred to as "a tentative division of the Lower Monroe."[1] The writer is convinced that the Tymochtee should be recognized as a distinct division, even if its limits are not now definitely known. At all places where the Niagaran-Monroe contact is shown, it is the typical Greenfield that is present. At no place is the Tymochtee known resting on the Niagaran or so near the Niagaran as to preclude a thickness of 50 feet or more of Greenfield from intervening. It may therefore be considered positive that the Greenfield is everywhere the basal member of the Monroe.

It is believed that there are all gradations from the typical Greenfield to the most shaly phase of the Tymochtee and that probably there is nowhere a sharp contact plane between typical Greenfield and typical Tymochtee. It is believed that the Greenfield type of deposit was laid down everywhere over the Monroe area to a depth of 50–100 feet and that then there was a relative increase of the argillaceous and carbonaceous material over the calcareous material gradually, or irregularly and unequally. It is probable that this transition did not take place everywhere at just the same time and that

[1] *Bull. Geol. Soc. Amer*, Vol. XIX, p. 554.

the intermediate type of deposit continued at some places much longer than at other places, but in time the typical Tymochtee deposits were being formed over the entire area of deposition but with unequal clay content at different places in conformity with nearness to the source of the clay, transporting agencies, etc.

The relation of the Tymochtee to the overlying strata is equally obscure in the central part of the state, but in Lucas and Ottawa counties the very shaly strata assigned to the Tymochtee are directly overlaid by the Put-in-Bay member.

The fauna of the Tymochtee is a very meager one in number of species and individuals. The two most usual fossils are a small *Hindella* and a small *Leperditia* which cannot be distinguished from *Hindella rostralis* and *Leperditia ohioensis* of the Greenfield except by their smaller size. The fauna is not distinct from that of the Greenfield and contains only a few depauperate forms which were able to live on in spite of the adverse conditions.

PUT-IN-BAY DOLOMITE

The name Put-in-Bay dolomite was proposed in 1907[1] for the rock division forming the prominent cliffs of Put-in-Bay (South Bass Island) in Lake Erie. The type of stone forming the Put-in-Bay member was recognized, however, in 1874 by Winchell, who described it in the Ottawa County report as "Phase No. 1" of the Waterlime[2] and gave for "its typical exposure" the bluffs about the south end of Put-in-Bay Island.

The Put-in-Bay member is most typically a gray-drab, brecciated, rough-textured, massive dolomite that weathers with irregular knobby surface, but the member contains some bedded stone in layers of 2–6 inches, and it was apparently in this form that all the stone was laid down. The member is best exposed in the cliffs on the northwest part of Catawba Island and around South Bass Island. At both places the exposure shows about 50 feet, which is apparently the practically complete thickness of the member, although the upper contact is not shown. In both of these exposures there are commonly two horizons of breccia separated by bedded strata. The upper zone of breccia is commonly about 25 feet thick, the lower zone about 10

[1] *Ibid.*

[2] *Geol. Surv. Ohio*, Vol. II, Part 1, pp. 230–31.

feet thick, and the intervening bedded stone from 5 to 10 feet thick. There are, however, many lateral changes from brecciated to bedded stone and vice versa, so that the thicknesses of the brecciated and bedded units vary from place to place, and locally practically the whole thickness of the member is brecciated. The lateral changes are commonly quite abrupt, but the type of stone forming the fragments of the breccia is the same as that forming the bedded stone a short distance laterally. Further, there is a direct relation between the thickness of the layers of bedded stone and the size of the fragments in the breccia resulting from that stone. The stone in layers of 3–8 inches formed a coarse breccia with cobble-like masses up to 4–6 inches across, while a zone of thin-bedded, laminated, shaly dolomite at the base of the Put-in-Bay member formed a breccia of small angular fragments 1 inch or less across.

By what process or processes the breccia was formed from the bedded stone is not known. It appears that the change took place soon after the sediments were laid down and before they were completely cemented to a firm stone. For the coarser type of breccia it is probable that the partly consolidated sediments were plowed up and broken up by great storm waves and that the fragments settled back with the interstices filled in with finer material. The whole mass was then compacted and cemented.

On Figure 1 the Put-in-Bay member is joined with the Raisin River. The Put-in-Bay has its chief distribution on the Bass islands and on Catawba Island, where it forms steep cliffs and over the higher parts of the islands forms a characteristic sink topography. It is known to the south on the axis of the Marblehead peninsula and at one place just south of Sandusky Bay at Crystal Rock Hill. No exposures of this member are known farther south along the line of outcrop, but this is a region of thick drift and no exposures showing its horizon are known. The Put-in-Bay member is known at but one place on the west side of the anticline. This is at the Holland Quarry in Lucas County, where are exposed about 34 feet of drab, firm, bedded dolomite and massive, brecciated dolomite which replace each other laterally and vertically and look very much like the Put-in-Bay dolomite of Ottawa County. This correlation is further supported by the finding in the lower few feet of the division of a few

specimens of *Leperditia alta* and a single specimen of *Eurypterus* sp., characteristic fossils of the Put-in-Bay. The member is not known anywhere farther south on the west side or over the axis of the anticline. It is apparently not present in Logan County, although no exposure covering its horizon has been seen.

It was upon the evidence furnished by the presence of *Leperditia alta*, *Eurypterus eriensis*, and *Spirifer ohioensis* found in the rocks of South Bass Island that Newberry first differentiated the Waterlime group in Ohio.[1] Grabau and Sherzer listed[2] a fauna of four species from the Put-in-Bay dolomite, consisting of *Spirifer ohioensis*, *Goniophora dubia*, *Leperditia alta*, and *Eurypterus eriensis*, and made *Goniophora dubia* the type fossil of the member. Various collections of fossils have been made by the writer from the Put-in-Bay exposures. The four species noted above are most common and most characteristic, but *Spirorbis laxus*, *Pterinea lanii*, and *Schuchertella hydraulica*, listed by Grabau as Raisin River forms, have been found in the Put-in-Bay.

RAISIN RIVER DOLOMITE

The name Raisin River was given in the classification of 1907[3] for the highest division of the Bass Island, and the type exposures may be considered as those along the Raisin River at Monroe, Michigan, and in the quarry 1 mile south. The Raisin River is commonly blue-gray to drab, banded, argillaceous dolomite with carbonaceous partings along the bedding planes. It is commonly in thin beds of 2–6 inches, but locally there are thick beds or massive ledges. The member contains layers or zones of mottled, speckled, and streaked dolomite which is very dense and compact and breaks with a conchoidal fracture.

The Raisin River member has its greatest development in Monroe County of southeastern Michigan, where it is reported to have a thickness of about 200 feet.[4] It is present in northern Ohio on both sides of the Cincinnati anticline. On the west side the best outcrop, and in fact the only complete section of the Raisin River known in

[1] *Op. cit.*, p. 16.

[2] *Op. cit.*, pp. 33–35.

[3] *Bull. Geol. Soc. Amer.*, Vol. XIX, p. 554.

[4] *Mich. Geol. and Biol. Surv.*, Pub. 2 (1910), p. 35.

the state, is at the Holland Quarry in Lucas County, but instead of the 200 feet reported for Monroe County, the member here is only about 50 feet thick. It is underlaid by the Put-in-Bay member and overlaid disconformably by the Sylvania sandstone. This is the only place at which a positive Raisin River fauna has been found on the west side of the anticline, but strata of the proper lithologic appearance and stratigraphic position for the Raisin River member were seen at several other places in Lucas County. No exposures of the Raisin River are known south of the Maumee River.

On the east side of the anticline the Raisin River member is known positively only on the Bass islands in Lake Erie, where there are a number of small exposures with the characteristic Raisin River fossils, all in the basal part of the member. No exposures of the Raisin River are known on the mainland south of the Bass islands, but the line of outcrop southward across Sandusky and Seneca counties is deeply drift covered.

Fossils are never abundant and rarely common in the Put-in-Bay. From the faunas collected, thirteen forms have been identified, several of which are without specific designation because of the poor character of the material. The more abundant forms are *Whitfieldella prosseri*, *Spirorbis laxus*, and *Pterinea lanii*, the first of which is, so far as known, limited to the Raisin River member. The other two species mentioned were restricted to the Raisin River by Grabau,[1] but both have been found in Ohio in other members of the Bass Island. Grabau listed a fauna of twenty species from the Raisin River member and made *Whitfieldella prosseri* its most characteristic fossil. All the common forms as found in Ohio are in Grabau's list. The only notable additions from Ohio are three corals and a stromatoporoid, there being no representatives of these in Grabau's list.

PUT-IN-BAY AND RAISIN RIVER IN THE CENTRAL PART OF THE STATE

No positively identified Put-in-Bay or Raisin River strata are known anywhere in Ohio south of Lucas County on the west side of the anticline or south of Erie County on the east side of the anticline. At a quarry near the west edge of Bellevue in the southeast corner of Sandusky County the Columbus limestone rests upon the Lucas

[1] *Ibid.*, pp. 211–13.

member with apparent conformity. Southward from here not a single exposure is known showing this contact until Delaware County is reached, where the Columbus lies upon the Bass Island formation. Exposures of the bordering Niagaran and Columbus limestones are such as to show that the Monroe belt is not very wide in Seneca and Wyandot counties, and therefore, since the dip is slight, that the thickness of the Monroe cannot be very great, and it is probable that the Detroit River members present in Erie County thin out southward in northern Seneca County and are entirely absent farther south.

The Monroe is exposed at several places along or near the Scioto River in western Delaware County and at a few places in Franklin and Fayette counties farther south, most of the exposures being at or near the contact with the Columbus limestone above. The dominant type of rock is gray, argillaceous, fine-grained, laminated dolomite, in beds of 2–6 inches, which weathers to a soft chalky stone of ashen-gray color. With this are associated some rougher-textured, darker layers with open spaces or with quartz-filled spaces, and some layers of brown or drab, oölitic dolomite. At several places fossils have been found which show that the rock is in the Bass Island formation, but they do not determine which member, although indicating the upper part of the formation rather than the Greenfield member at the base.

Whitfield reported the presence of *Goniophora dubia* in the Lower Helderberg group at Middletown (now called Prospect), Marion County.[1] This fossil was later designated by Grabau as the type fossil of the Put-in-Bay member, and apparently upon its reported presence at Middletown by Whitfield, Grabau stated that the Put-in-Bay member was present at Middletown (Prospect).[2] *Goniophora dubia* has however also been found in the Raisin River member. The presence of the Put-in-Bay member is indicated in Franklin County by the statement by Stauffer that *Spirifer ohioensis*, which is limited to the Put-in-Bay member by Grabau and Sherzer[3], has been found in the Monroe of the Columbus quadrangle.[4] Neither of the

[1] *Geol. Surv. Ohio*, Vol. VII (1893), p. 416.

[2] *Bull. Geol. Soc. Amer.*, Vol. XIX (1907), p. 554.

[3] *Op. cit.*, pp. 136 and 211.

[4] *Geol. Surv. Ohio*, Bull. 14 (1911), p. 18.

foregoing records has been verified by the writer, by finding again the fossils indicated, although careful search has been made in both regions.

In general lithological characteristics the Monroe strata of western Delaware and Franklin counties suggest the Raisin River rather than the Put-in-Bay member. Faunally, however, the evidence is against this correlation, for the diagnostic fossil of the Raisin River, *Whitfieldella prosseri*, is not present, while *Leperditia*, unknown in the Raisin River of northern Ohio, is present here wherever any fossils were found. Farther westward in Union County there are a few outcrops of the Tymochtee member, scattered over a large area indicating a considerable thickness of strata, which, on the basis of the general eastward dip of the rocks of this part of the state, should pass beneath that part of the Monroe exposed near the Columbus boundary in western Delaware County. Well records in Delaware and Franklin counties are interpreted as showing 300–400 feet of the Monroe, and the strata exposed along the Columbus boundary should be at the top of this thickness.

On the basis of the data now available it can only be stated that the Monroe which outcrops near the Columbus-Monroe boundary in western Delaware and Franklin counties is in the Bass Island formation above the Tymochtee member and probably below the Raisin River member, in the general horizon of the Put-in-Bay member 300–350 feet above the base of the Monroe. However, the rock is not brecciated like the typical Put-in-Bay dolomite of northern Ohio, and the characteristic Put-in-Bay fossils have not been found, although such meager faunal evidence as exists favors this member. It is probable that the fossils used to distinguish the Put-in-Bay and Raisin River members in northern Ohio do not exist or do not show the same associations here, and the exposures of the region are so few and scattered that it has not been possible as yet to establish different members if such exist.

SYLVANIA SANDSTONE

The rock unit which we now know as the "Sylvania sandstone" was first differentiated by Newberry,[1] who called it the "Oriskany

[1] *Op. cit.*, p. 17.

sandstone" from the sandstone at the same general horizon in New York. In 1888, Dr. Edward Orton showed[1] that this sandstone of Lucas County was not at the base of the Corniferous (Columbus) as the Oriskany should be, but that about 200 feet of "unmistakable Waterline" intervened between it and the base of the Corniferous. To this sandstone Orton gave the name Sylvania, and included it in the Waterline division. In the classification proposed in 1907 it was made the middle unit of a threefold subdivision of the Monroe.[2]

The Sylvania sandstone is an even-grained, quartz sandstone consisting of well-rounded grains loosely cemented. It has its best development and greatest thickness in southeastern Michigan, and in that region has been interpreted by Grabau and Sherzer as an eolian deposit.[3] In Ohio it outcrops in a narrow north-south belt from the Michigan state line south across Lucas County (Fig. 1). The region is drift covered, and only four or five exposures are known in this distance. South of where it crosses the Maumee River no exposures are known, nor is it known to exist elsewhere in Ohio.

The Sylvania sandstone rests disconformably upon the Raisin River below. Only a few exposures of this contact have been found, but they all show the disconformity. The contact surface is slightly uneven, with the sandstone fitting down over the truncated layers of the Raisin River dolomite. Further, the sandstone has a basal conglomerate the pebbles of which came from the underlying dolomite. The Sylvania sandstone is conformable with the overlying dolomite, which in this region is the highest beds of the Amherstburg, the lower part of the Detroit River being absent. The sandstone grades upward into dolomitic sandstone; then into arenaceous dolomite; then into dolomite; not as a uniformly gradual change, but with alternating layers of more sandy and less sandy material with a general decrease upward of the sand content. Stratigraphically, therefore, the Sylvania of this region is closely related to the overlying Detroit River formation and distinctly set off from the Raisin River member of the Bass Island formation below.

Fossils exist in the arenaceous dolomites that form the transition from the Sylvania to the Amherstburg, and in fact a few fossils have

[1] *Op. cit.*, pp. 18–20.

[2] *Bull. Geol. Soc. Amer.*, Vol. XIX, pp. 553–56.

[3] *Op. cit.*, pp. 61–86.

been found in the Sylvania sandstone down to within a foot of its base. The fossils found in the Sylvania, and in the transition beds above, are the same as in the overlying member of the Detroit River and are not related in the least to those of the underlying Bass Island.

The eolian origin of the Sylvania sandstone, as used by Garbau and Sherzer for southeastern Michigan, cannot be applied to most of the Sylvania of northwestern Ohio, where there is much less of the pronounced cross-bedding and most of the sandstone is definitely bedded as an aqueous deposit. Further, most of the sandstone is more or less dolomitic, especially in the upper part, and the fossils found in the Sylvania and in the transition beds are marine fossils.

The Sylvania sandstone, as found in Ohio, is interpreted as being largely water laid under marine conditions, although the wear of the grains may have been the result of previous windwork. The on-coming sea of the Detroit River stage, which reached northwestern Ohio in late Amherstburg or early Lucas time, probably advanced over a region of thin eolian sand which it re-worked and redeposited as an aqueous deposit, in which were incorporated remains of the life existing in this encroaching sea. The Sylvania sandstone is not of the same age at all places. In its water-laid form it is the earliest deposit of the on-coming Detroit River sea and is much later in this region than the Sylvania sandstone of the region farther north, where it is overlaid by the Flat Rock member. As an aqueous deposit it is here of late Amherstburg or early Lucas age.

Stratigraphically and faunally, the Sylvania is so closely related to the overlying member that to whatever age the Detroit River is assigned, the Sylvania must go with it.

DETROIT RIVER FORMATION

In 1903, Prosser proposed to call the Monroe strata above the Sylvania sandstone in northwestern Ohio the "Lucas limestone."[1] Shortly thereafter, work in southeastern Michigan and southwestern Ontario showed that a much more extended section of the Monroe above the Sylvania sandstone exists in the Detroit River region, and the classification of the Monroe put out in 1907 gave the name Detroit

[1] *Op. cit.*, p. 540.

River for that part of the Monroe above the Sylvania sandstone,[1] and divided it into four members as shown in Table I. The two lowest members, the Flat Rock and the Anderdon, have not been recognized in Ohio. They are known to be missing on the west side of the anticline in Lucas County, where their horizon is exposed, and although their horizon is not exposed on the east side of the anticline in Ottawa, Erie, and Sandusky counties, it is believed that they are absent here also. They are certainly absent in the central part of the state in Delaware and Franklin counties. The two upper members of the Detroit River formation, the Amherstburg dolomite and the Lucas dolomite, are present on the outcrop in Ohio. They are here discussed together because they are commonly found in the same outcrops and because at places it is not possible to fix exactly the boundary between them.

AMHERSTBURG AND LUCAS DOLOMITES

The Amherstburg dolomite takes its name from the village of Amherstburg on the east bank of the Detroit River in Ontario, near which place it was first studied in channel-improvement exposures in the Detroit River. It has been recognized also in a few quarries in Monroe County, Michigan,[2] and in the shaft of the Oakwood salt mine near Detroit, where it has a thickness of about 20 feet.[3] The name "Lucas limestone" was proposed in 1903 by Prosser for all the Monroe of northwestern Ohio above the Sylvania sandstone,[4] and in the classification of the Monroe put out in 1907 the term Lucas was redefined as the highest of the four divisions of the Upper Monroe.[5] This was practically what Prosser intended that the name should cover, for the two lowest divisions of the Detroit River are absent in Lucas County, and the third is very poorly developed.

The Amherstburg is commonly drab or brown dolomite, chiefly in thick layers some of which are porous or have small cavities. As a whole the strata of this member are browner, thicker bedded, and more porous than in any other member of the Monroe. The Lucas

[1] *Bull. Geol. Soc. Amer.*, Vol. XIX, pp. 553–56.

[2] *Mich. Geol. and Biol. Surv.*, Pub. 2 (1910), pp. 48–51.

[3] *U.S. Geol. Surv., Folio No. 205* (1917), p. 19, Fig. 19.

[4] *Op. cit.*, p. 540.

[5] *Bull. Geol. Soc. Amer.*, Vol. XIX, p. 556.

is gray to drab, bedded dolomite commonly in layers of 3–12 inches and with carbonaceous partings between the layers. There are also some thicker, rougher-textured layers and some denser, mottled stone.

The Amherstburg fauna is the largest fauna of the Monroe, more than fifty species being known. Most of the forms are larger than is usual in the other members, and it is therefore the most robust fauna of the Monroe. Many of the forms have distinct Devonian relations, and this member offers the strongest evidence for the Devonian age of the Detroit River formation. A distinctive element of the Amherstburg fauna is an abundance of corals, especially *Cladopora bifurcata*. However, practically all of the corals have been found also in the Lucas. Aside from the corals, the characteristic Amherstburg species are rare, but most of the collections contain a few specimens of one or more of the following species which are known only in the Amherstburg or at least not in other members of the Detroit River: *Fenestella* sp., *Schuchertella amherstburgensis*, *Strophеodonta vasculosa*, *Strophеodonta demissa* mut. *homolostriata*, *Spirifer sulcata* mut. *submersa*, *Dawsonoceras annulatum* var. *americanum*, *Cyrtoceras orodes*, and *Proetus crassimarginatus*.

The Lucas fauna includes about forty species, and, in contrast with the Amherstburg, the shells are commonly small. The most usual assemblage includes several species of minute gastropods of the genera *Acanthonema*, *Hormotoma*, and *Solenospira*, two or more species of the brachiopod genus *Prosserella*, including *Prosserella lucasi*, and one or two species of cylindrical corals.

Grabau's check list gives about thirty species restricted to the Amherstburg and about twenty species restricted to the Lucas. The work in Ohio has resulted in the extension of the range of a number of Grabau's Amherstburg species into the Lucas and the extension of the range of a number of his Lucas species into the Amherstburg. This is particularly true among the corals. Most of the species known only from one member are rare forms, and most of the common forms range through both members. It is not therefore always possible to distinguish between these members on the basis of a small fauna. Larger or several faunas commonly bring out the separation.

Amherstburg fossils are present in Lucas County in the water-

laid phase of the Sylvania sandstone and in the arenaceous dolomite strata that form the transition to the overlying Lucas dolomite. There is no massive, brown dolomite here as typical of the Amherstburg elsewhere, and the Amherstburg time closed before the transition to the true dolomite was completed.

The Lucas is the chief division of the Upper Monroe of northwestern Ohio, and its outcrop forms the west edge of the Monroe area across Lucas County. It is exposed in several quarries and on the Maumee River bed at several places between Otsego Rapids and Grand Rapids. No exposures of the Detroit River formation are known southwest of the crossing of the Maumee River. The greatest exposure is in the quarry of the Toledo Stone Company at Silica, where the member is about 140 feet thick and all exposed except about 10 feet at the top. The stone is drab dolomite with some banding and lamination and with some layers which are streaked and mottled with blue. The Lucas dolomite is overlaid by the Columbus limestone, the actual contact being exposed at only one place, which is in the Maumee River Valley at Grand Rapids, where it shows a disconformable plane with a relief of at least a few feet. The lithologic contrast of the two formations is prominent, and the contact plane is locally a prominent stylolite suture.

On the east side of the anticline the Detroit River formation is known in Ottawa, Erie, and Sandusky counties. There are exposures around the west shore line of Kelley's Island (Lucas); on the north shore of Marblehead Peninsula west of Marblehead (Amherstburg and Lucas); over an extent of several square miles east of Castalia (Lucas); at the quarry of the Wilson Stone Company $1\frac{1}{2}$ miles southwest of Castalia (Amherstburg and Lucas); and in a quarry near the west edge of Bellevue in the southeast corner of Sandusky County (Lucas).

The best exposures are along the lake shore on the north of Marblehead Peninsula in Ottawa County. The exposures are along the shore line or in low cliffs rising 5–10 feet above the water, but the strata dip slightly to the east and a considerable section is exposed. East of the village of Marblehead the Columbus limestone forms the shore, but going west from this village the base of the Columbus rises, and in the next $1\frac{1}{2}$ miles through Lakeside successively lower

zones through the Lucas and the Amherstburg appear above the water to a total thickness of about 80 feet. The highest 30–35 feet of these strata are of Lucas age, although their fossils here are not very distinctive. The Lucas strata are thinner bedded and more even grained than the Amherstburg below. The highest strata interpreted as Amherstburg appear about ½ mile west of the Life Guard Station at Marblehead, and from here westward through Lakeside for more than a mile successively lower zones are exposed to a thickness of about 50 feet of Amherstburg without reaching its base. In the upper part there are several massive ledges of brown, porous dolomite quite like the stone from the Detroit River channel at Amherstburg. This is the best exposure of Amherstburg in Ohio.

The Lakeside exposure terminates westward in beach deposits without reaching the base of the Amherstburg. Two and one-half miles farther west the Amherstburg is again exposed in the low elevation forming the southeast part of Catawba Island peninsula, where 10–15 feet of rock outcrop. If the usual eastward dip of the region holds between this exposure and the Lakeside exposure, these strata should be 75–100 feet below those at the west end of the Lakeside exposure, thus requiring a thickness of about 150 feet of Amherstburg. This thickness is improbable, and the rocks probably lie essentially horizontal in this unexposed interval. There are, however, certainly 50–75 feet of Amherstburg strata here in Ottawa County, which is much greater than known elsewhere, and the total thickness may be 150–200 feet.

Twelve miles south of Lakeside, at the quarry of the Wilson Stone Company 1½ miles southwest of Castalia, there is exposed a section extending down from the Columbus limestone 50 feet into the Detroit River formation. Only a few fossils were found in the Detroit River strata here, and the plane of separation of the Lucas and Amherstburg is not very definite, but on the basis of lithology and such fossils as exist it is believed that about 30 feet of Lucas are present—a thickness similar to that on the north shore of the Marblehead Peninsula. The lower 20 feet of the exposure at this quarry are interpreted as Amherstburg. The stone is thicker bedded than the Lucas above and contains a few of the characteristic Amherstburg fossils. Just east and northeast of Castalia small exposures

were seen from which good Amherstburg faunas were collected. The faunas collected here and at Lakeside contain a number of the distinctive Amherstburg species, such as *Proetus crassimarginatus*, *Schuchertella amherstburgensis*, *Spirifer sulcatus* mut. *submersa*, etc. There are also present here several new forms, not known elsewhere, which have their closest relationships with early and middle Devonian species of New York.

The Columbus-Lucas contact was seen at several places on the east side of the anticline, and at all these places it is apparently conformable. Commonly the lithologic and faunal change takes place at a bedding plane no more prominent than others above and below, and at several places the contact is within a ledge. The Columbus forms of life appeared suddenly and the few Lucas forms existing died out suddenly. At one place there are a few Lucas fossils in the basal 2 inches of the basal layer of Columbus limestone.

Farther south along the east side of the anticline no outcrops of the Detroit River formation are known. The formation is absent in central Ohio, in Delaware and Franklin counties, and it is probable that the members which exist in Erie County thin out southward or are overlapped by the Columbus limestone in northern Seneca County.

The Detroit River formation is present around the base of the Bellefontaine outlier on the crest of the anticline in Logan County. Several exposures of the Lucas dolomite are known, and one exposure exists that is probably Amherstburg.

THE FAUNAS OF THE MONROE

Practically all the fossils previously reported from rocks which we now call Monroe in Ohio were collected from either the Greenfield or the Put-in-Bay members and were recognized as related to Upper Silurian fossils of other regions. On this basis the Monroe has been interpreted in Ohio as Silurian.

The work of Grabau and others in the region of the Detroit River first demonstrated the extent of the section and the succession of the members of the Monroe above the Sylvania sandstone. In the study of the faunas of these Upper Monroe or Detroit River members, Grabau recognized their distinctness from the faunas of the

Bass Island formation and the many Devonian aspects of the faunas of the three lower members of the Detroit River.[1] However, the position of these three members beneath the Lucas member, containing a fauna which Grabau emphasized strongly as Silurian, caused him to conclude that the whole of the Detroit River formation was Silurian.

In 1916, Stauffer showed the close relationship faunally of the Amherstburg to the Onondaga and suggested that a place for the Detroit River might be found in an early Devonian northwest embayment in part contemporaneous with the Helderbergian of the Atlantic embayment.[2] In connection with a study of the "Silurian of the Ontario Peninsula"[3] Williams in 1919 made a reappraisement of the faunal evidence as to the age of the Detroit River formation and concluded that it is Devonian.

The study of the faunas of the Bass Island and Detroit River formations in Ohio has shown more contrast than similarity, only a few long-ranging, indefinite species being common to the two faunas. The Bass Island fauna is distinctly Silurian. The Detroit River fauna has many Devonian aspects, and especially is this true of the Amherstburg fauna. There are more related forms in the Detroit River dolomite and the overlying Columbus limestone than in the two formations of the Monroe. It is not possible here to enter into an analysis of the Detroit River faunas as to their Devonian or Silurian age. Such analyses have been given by Grabau, Stauffer, and Williams with the divergent conclusions noted above, and an analysis will be made, in the light of such new evidence as has appeared in the Ohio study, in a later complete report on this work. The evidence is very conclusive for the Devonian age of the Detroit River formation.

SUMMARY OF DISTRIBUTION OF MEMBERS AND INTERPRETATION OF THE GEOLOGIC HISTORY

The extent of the Monroe in Ohio is shown by the total shaded area of Figure 1. Most of this area belongs to the Bass Island formation, the Detroit River formation being present only in narrow bands on either side of the anticline in the north part of the state and around the Bellefontaine outlier in west-central Ohio. The Green-

[1] *Op. cit.*, pp. 215–23. [2] *Loc. cit.* [3] *Loc. cit.*

field and the Tymochtee have the greatest areal and actual distribution (Fig. 1). The Greenfield is the only member present in the extreme southern part of the area and the only one which extends across the state from north to south. It is apparently present beneath all other members, and its actual distribution therefore includes the entire Monroe area. It is present on either side and over the axis of the anticline, and its deposition was apparently not influenced in any way by this structural feature. The sea in which it was deposited extended from Lake Erie to the Ohio River, covered most of northwestern Ohio, and may have covered considerable area in southwestern Ohio from which the rocks of this horizon have been eroded. It also covered an unknown area to the east, north, and northwest where the Monroe passes beneath younger formations.

The Tymochtee shaly dolomite has its greatest development over the axis of the anticline in the middle part of the Monroe area, but it is present also on either side of the anticline in northern Ohio, at places over the west prong of the Monroe area, and southward to southeastern Fayette County. It is present in all parts of the region where its horizon appears on the outcrop, and these areas indicate an original distribution probably as great as that for the Greenfield. Its less resistant character, however, has allowed its removal from a large area over the northern part of the anticline, thus exposing the Greenfield below.

The Put-in-Bay and the Raisin River have been positively differentiated only in northern Ohio, in Lucas County on the west of the anticline, and in Ottawa and Erie counties on the east. Both of these members appear to thin southward in Lucas County, and they are not known anywhere over the west prong of the Monroe area or in the Bellefontaine outlier. However, strata belonging to the Bass Island formation, apparently above the Tymochtee, exist along the east edge of the Monroe area in central Ohio, in Delaware, Franklin and Fayette counties, but they have not yielded the typical fossils of the Put-in-Bay or the Raisin River. Toward the close of the Bass Island the sea withdrew entirely from Ohio, probably to the north, and there followed a time of erosion in northwestern Ohio, as shown by the disconformity at the top of the Bass Island formation. To the northwest in Michigan, extensive and thick deposits of eolian

sand were laid down at this time, and the thinning edge of this area extended into Lucas County.

There next followed the invasion by the Detroit River sea. This sea came in from the north, for the two basal members of the Detroit River formation, the Flat Rock and the Anderdon, which are present in the Detroit River region, are not present in Ohio. The sea reached Ottawa and Erie counties in early Amherstburg time, and there was formed the 100 feet or more of Amherstburg dolomite present there. The sea was probably encroaching westward and southward, for, toward the close of the Amherstburg time, it reached Lucas County. Here there rested upon at least part of the region the eolian sand deposit noted above. This sand was re-worked by the on-coming sea, shifted more or less and spread out more widely, especially to the south. As redeposited it was a water-laid sediment mixed with more or less dolomitic material, grading upward into the true dolomite, and contained the remains of the Amherstburg forms of life living in this sea. Near the state line, where the eolian sand was probably thickest, the re-working did not extend to its base, and in part its eolian structures are still preserved as shown in the quarries at Silica. Farther south the entire deposit was apparently re-worked, for we find Amherstburg fossils down to the base of the Sylvania at the Holland quarry. It is just the closing phase of the Amherstburg that is represented by the re-worked sandstone and the transition beds above in Lucas County, and the sea did not reach the south line of the county until after the Lucas fauna had come in.

Sometime after the erosion of the emerged surface of the Bass Island formation and before the re-working and redeposition of the Sylvania sandstone in late Amherstburg time, there was deposited in one of the small irregularities cut into the Raisin River member at the top of the Bass Island formation a black carbonaceous shale containing remains of ostracoderms of genera known elsewhere only from early Devonian deposits. This pocket of black shale is interpreted as a fresh-water deposit in a stream channel or other basin.

Deposition continued in northern Ohio through Lucas time, but whereas the Amherstburg accumulation was greatest on the east in Ottawa County, the Lucas accumulation was greatest on the west in Lucas County, and an arm of the sea extended southward along what

is now the axis of the anticline to beyond Logan County, where the Lucas is well developed in the Bellefontaine outlier, but apparently did not cover Delaware and Franklin counties to the east. The distribution of the Detroit River members indicates a gradually extending sea southward, which reached its greatest extent in this Lucas sea.

At the close of the Detroit River time the sea withdrew from the southward extension to Logan County and from Lucas County, for in both these regions the contact with the overlying Columbus limestone is disconformable. The recession was to the northeast and was not complete, for all the evidence furnished by the contact exposures in Ottawa, Erie, and eastern Sandusky counties indicates that here deposition continued unbroken from the Lucas into the Columbus time.

SUMMARY OF POINTS BEARING ON THE AGE OF THE MONROE

In the foregoing pages the following points bearing on the age of the Monroe have been emphasized: (1) The geographical distribution of the several members of the Monroe shows that in Ohio there was an early Monroe (Bass Island) sea advance and a late Monroe (Detroit River) sea advance separated by a complete withdrawal of the sea in Middle Monroe (Sylvania) time. (2) The faunas of the Bass Island and Detroit River formations have very little in common. The Bass Island fauna is definitely Silurian. The Detroit River fauna is better interpreted as Devonian. (3) The Sylvania sandstone is stratigraphically and faunally so closely related to the overlying Detroit River formation that whatever age is assigned to the Detroit River, the Sylvania sandstone must go with it. (4) A pronounced stratigraphic break (disconformity) and a distinct faunal change are located at the base of the Sylvania sandstone. (5) A small pocket of shale resting on the Bass Island and beneath the Sylvania sandstone contains distinctive fossil ostracoderms known elsewhere in the world only in the basal Devonian.

On the basis of the foregoing lines of evidence it is concluded that the Detroit River dolomite and the Sylvania sandstone beneath it are Lower Devonian in age, and that the Silurian-Devonian contact in Ohio is at the base of the Sylvania sandstone.

TWIN EARTHQUAKES

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ABSTRACT

As a rule, the earthquakes of Great Britain are weak and infrequent. Once a month, on an average, the ground is shaken somewhere within its limits. Once in every twelve years, a shock occurs that is strong enough to throw down chimneys or fracture walls. Though simple in nature and of brief duration, there are, however, two well-marked types of earthquakes, which I have called "simple" and "twin" earthquakes. In the first section of this paper, I propose to point out the differences between the two classes; in the second, to describe some typical twin earthquakes; in the third, to state the characteristic features of twin earthquakes in general; in the fourth, to show that twin earthquakes originate in two detached foci; in the fifth, to notice that there are two classes of twin earthquakes and to summarize their essential features and points of difference; and, in the sixth, to consider the origin of twin earthquakes and to mention some further properties of these earthquakes.[1]

DIFFERENCES BETWEEN SIMPLE AND TWIN EARTHQUAKES

Nature of the shock.—The most manifest difference between simple and twin earthquakes lies in the nature of the shock. In a simple earthquake, the vibrations increase in strength to a maximum and then die away, as at *a*, Figure 1. In a twin earthquake, the shock consists of two distinct parts, as at *b*, or of two parts connected by weaker tremulous motion, as at *c*. It will be seen afterward that these two types of vibrations, *b* and *c*, are characteristic of different classes of twin earthquakes. In both cases, the interval of rest between the two parts, or of tremulous motion between the two maxima, is brief. It is usually about 2 or 3 seconds.

Frequency and disturbed areas of principal shocks.—During the 28 years from 1889 to 1916, there were 11 twin earthquakes in Great

[1] Accounts of the twin earthquakes referred to in this paper are given in my *History of British Earthquakes* (Cambridge University Press, 1924), pp. 183–203, 206–12, 244–89, 337–44, with the exception of the Hereford earthquake of 1926 (*Geol. Mag.*, Vol. LXIV [1927], pp. 162–67). The original and fuller reports are given in my *Hereford Earthquake of December 17, 1896* (Cornish Brothers, Birmingham, 1899) and in the following journals: *Quar. Jour. Geol. Soc.*, Vol. LIII (1897), pp. 157–75; Vol. LVIII (1902), pp. 371–76; Vol. LX (1904), pp. 215–32; Vol. LXI (1905), pp. 1–7, 8–17; Vol. LXII (1906), pp. 5–12; Vol. LXIII (1907), pp. 351–61; *Geol. Mag.*, Vol. V (1908), pp. 301–3; Vol. VI (1919), pp. 302–12.

Britain, 2 of intensity 8, 6 of intensity 7, and 3 of intensity 5 (Rossi-Forel scale). In the same interval, there were 25 simple earthquakes of intensity above 4, 1 of intensity 8, 4 of intensity 7, 7 of intensity 6, and 13 of intensity 5. Thus, of the earthquakes of intensities 8 and 7, 62 per cent were twins; of those of intensities 6 and 5, 13 per cent were twins, and, of all those of intensities above 4, 31 per cent. Or, including all the earthquakes, 366 in number, during the same years, 11 were twins and there were 35 minor shocks associated with them; that is, 12 per cent of the total number were either twin earthquakes or their accessory shocks.

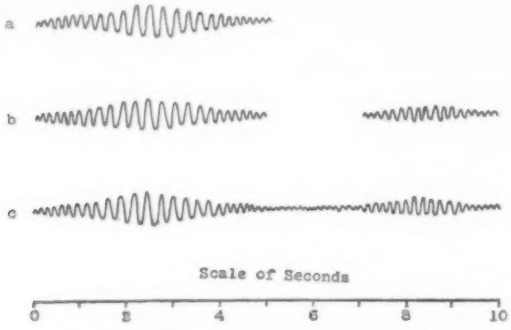

FIG. 1.—Nature of simple and twin earthquakes

In most of the earthquakes here considered, the disturbed area is bounded by an isoseismal of intensity 4; and, in estimating average disturbed areas, I have included only the area within this isoseismal. During the interval 1889–1916, the areas of 8 twin earthquakes of intensities 8 and 7 range from 10,120 to 98,000 square miles, with an average of 32,700 square miles. For 5 simple earthquakes of the same intensities the areas range from 605 to 33,000 square miles, with an average of 11,200 square miles. For 3 twin and 13 simple earthquakes of intensity 5 the average areas are 1,640 and 384 square miles. Lastly, for 11 twin and 25 simple earthquakes of intensities 8–5 the average areas are 24,300 and 2,680 square miles.

Thus, while twin earthquakes are much less frequent than simple earthquakes, in the ratio of about 1:8, the areas disturbed by them are much larger, in the ratio roughly of 3:1.

Frequency and disturbed areas of accessory shocks.—Omitting the slight twin earthquakes of Leicester in 1893, Carlisle in 1901, and Derby in 1906, the remaining 8 twin earthquakes (of intensities 8 and 7) were preceded by 12 foreshocks and followed by 19 aftershocks, or, with each earthquake, there were, on an average, 1.5 foreshocks and 2.4 aftershocks. With the 3 stronger simple earthquakes of the same intensities, there were 2 foreshocks and 34 aftershocks, or an average per earthquake of 0.7 foreshock and 11.3 aftershocks. Thus, the number of accessory shocks associated with a twin earthquake is to that associated with a simple earthquake as 1:3.

The disturbed areas (or areas within the isoseismal 4) are known of 14 accessory shocks of twin earthquakes and of 13 accessory shocks of simple earthquakes. The former areas range from 112 to 29,000 square miles, with an average of 3,630 square miles; the latter range from 35 to 2,200 square miles, with an average of 437 square miles. Some of the areas for the minor shocks of twin earthquakes are of remarkable size. For instance, one aftershock of the Pembroke earthquake of 1892, of intensity about 4, disturbed an area of about 4,800 square miles; and another, of intensity about 5, an area of about 29,000 square miles. Again, 2 foreshocks of the Hereford earthquake of 1896, both of intensity about 4, disturbed areas of about 6,300 and 6,400 square miles.

DESCRIPTION OF SOME TYPICAL TWIN EARTHQUAKES

Of the 12 twin earthquakes that I have studied, I select 4 for which the materials are unusually full: the Derby earthquake of March 24, 1903, the Stafford earthquake of January 14, 1916, the Doncaster earthquake of April 23, 1905, and the Leicester earthquake of August 4, 1893.

Derby earthquake of 1903.—On the map of this earthquake (Fig. 2), the continuous curves represent isoseismal lines of intensities 7 to 4, the area of the first being 112 square miles and of the last 12,000 square miles. The dotted lines are isacoustic lines of 95 and 90 per cent, the meaning of the former line being that, within a small area with its center on the line, 95 per cent of the observers heard the accompanying sound.

Throughout the whole of the disturbed area, with the exception

of a narrow band, the shock consisted of two distinct parts separated by an interval of 2 or 3 seconds of rest and quiet. So clear was this division of the shock that it was noticed by 68 per cent of all the observers. In places, the percentage rises to 80; in no large area does it fall below 48. At Ashbourne, the first part of the shock was the stronger; at Darley Dale, the second. At Derby, the interval be-

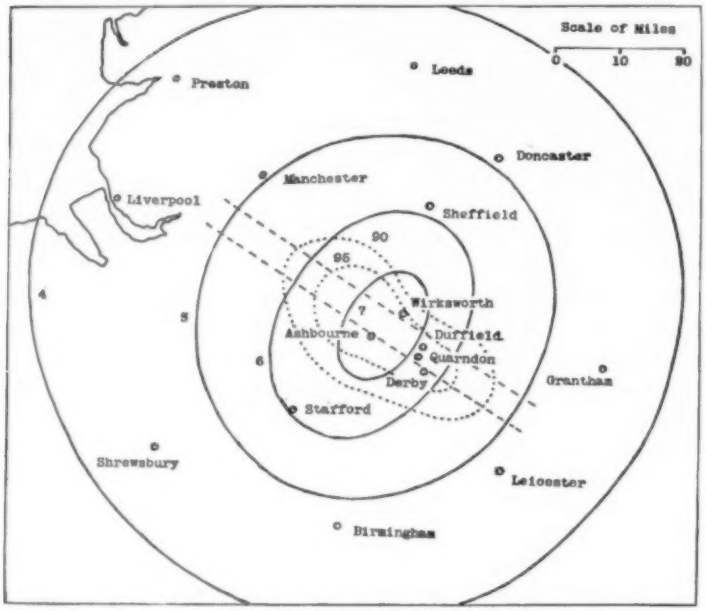

FIG. 2.—Derby earthquake of 1903

tween the two parts was no more than half a second. At Quarndon, the shock was a simple one with two maxima of intensity. At Duffield, only one shock with one maximum of intensity was felt.

The observations at Quarndon and Duffield are characteristic of all places within a rectilinear band, bounded on the map by broken lines. The band is about 5 miles wide and runs centrally across the inner isoseismals at right angles to their longer axes. I have called it the "synkinetic band," for, within it, the two movements were felt together.

Outside this band, the double shock was felt close to the boundary of the disturbed area; and this shows that the impulses producing the two parts were nearly equal in strength. This is also clear from the fact that, on the northeast side of the synkinetic band, 60 per cent of the observers regarded the first part as the stronger; and, on the southwest side, 63 per cent.

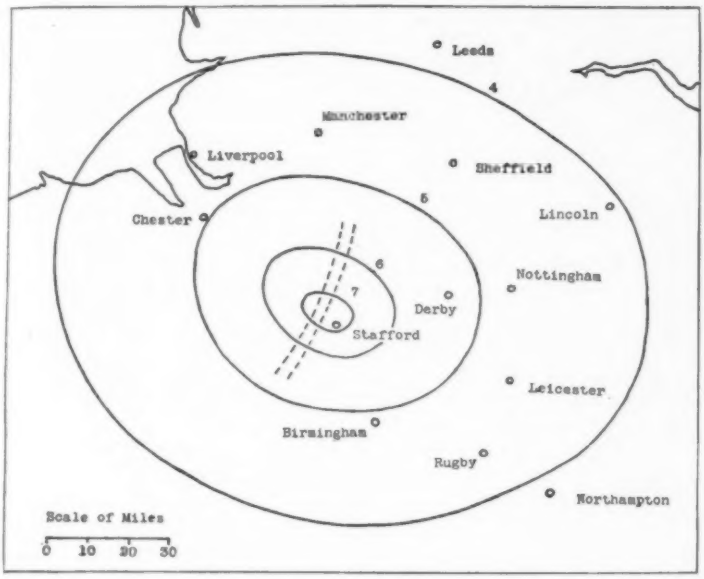

FIG. 3.—Stafford earthquake of 1916

Stafford earthquake of 1916.—On the map of this earthquake (Fig. 3), the continuous lines represent isoseismals of intensities 7 to 4 —the isoseismal 3, which bounds the disturbed area, being omitted. The area of the isoseismal 4 is 12,500 square miles; and the whole disturbed area, 50,200 square miles. The two broken lines represent the boundary of the synkinetic band; and it will be noticed that they are hyperbolic in form, with the concavity facing west.

As in the Derby earthquake, the shock shows every stage, from complete separation into two parts with an interval of about 2

seconds, to a single shock with two maxima just within the synkinetic band, and a single shock with one maximum at places near its axis.

Except within the synkinetic band, the double shock was felt over nearly all the disturbed area. The two parts of the shock were

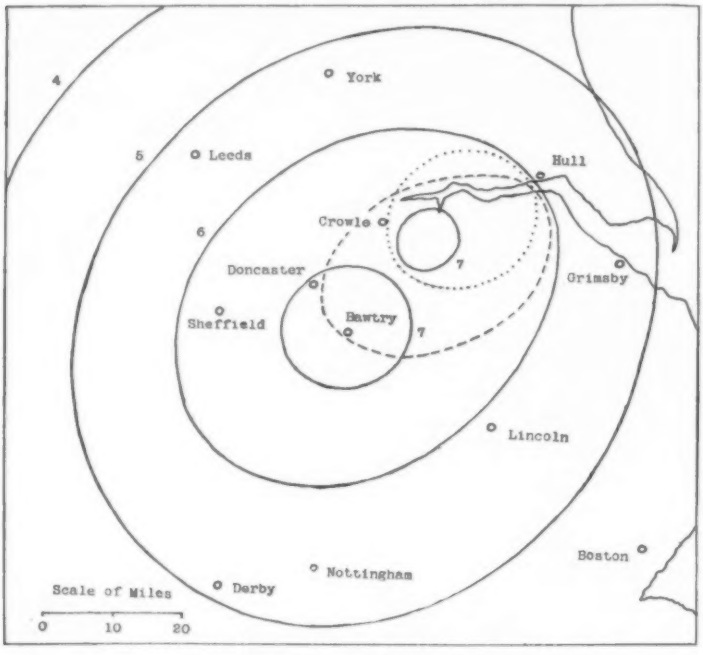

FIG. 4.—Doncaster earthquake of 1905

thus nearly equal in strength, but, on the east side of the band, the first part was slightly the stronger, and, on the west side, the second.

Doncaster earthquake of 1905.—On the map of this earthquake (Fig. 4), the continuous lines represent isoseismals of intensities 7 to 5, and the isoseismals 4 and 3 are omitted. The most interesting feature of the lines is the duplication of the innermost isoseismal, the area of the southwestern portion being 244 square miles and that of the other 63 square miles. The area included within the isoseismal

4 is about 10,700 square miles, and that within the isoseismal 3, about 17,000 square miles.

In this earthquake, the double shock was observed over an area that overlapped the isoseismal 5 in all directions and extended nearly to the isoseismal 4. Over the whole disturbed area, 32 per cent of the

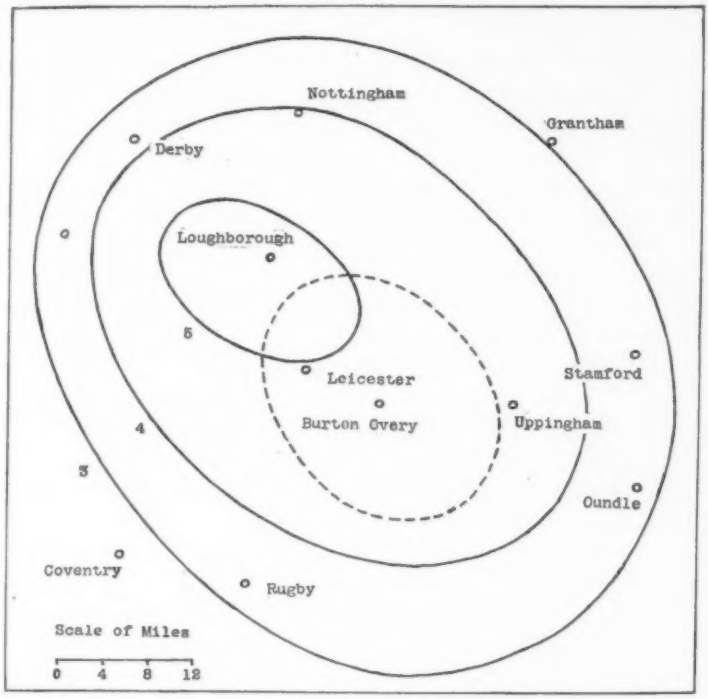

FIG. 5.—Leicester earthquake of 1893

observers noticed either two maxima in a continuous series of vibrations at places within and near the inner isoseismals or two detached series of vibrations, separated by an interval of a few seconds, at greater distances. The mean duration of the interval in such cases was 3.5 seconds. The two parts of the shock differed in strength, the first being slightly the stronger all over the disturbed area except within the small and nearly circular district indicated by the dotted

line in Figure 4. This area, it will be seen, includes the smaller and northeastern portion of the isoseismal 7, but its center lies about 6 miles to the northeast of the center of that portion. The broken line represents the isoseismal 4 of the Doncaster earthquake of April 13, 1902. This was a simple earthquake felt over a wide area, the isoseismal 4 containing about 600 square miles and the whole disturbed area about 3,000 square miles.

Leicester earthquake of 1893.—The continuous lines on the map (Fig. 5) represent the isoseismals 5, 4, and 3 of this earthquake; the broken line represents the isoseismal 5 of the earthquake of June 21, 1904. It will be noticed that the isoseismal 5 of the earlier earthquake is excentric with regard to the isoseismal 4, the distance between their centers being $8\frac{1}{2}$ miles.

Over nearly all the disturbed area, the shock consisted of two distinct parts separated by an interval of mean duration 2.5 seconds. The evidence as regards the relative intensity of the two parts is not very complete, but it suggests that the first part was the stronger all over the disturbed area.

CHARACTERISTIC FEATURES OF TWIN EARTHQUAKES

Wide area over which the twin shock is felt.—That the twin shock is no mere local phenomenon is clear from the wide area over which

TABLE I

Earthquake	Disturbed Area in Square Miles
Hereford, 1896	98,000
Carmarthen, 1893	63,600
Stafford, 1916	50,200
Pembroke, 1892	44,860
Derby, 1903	12,000
Leicester, 1893	2,200

it is observed. In the earthquake listed in Table I, it was felt over practically the whole disturbed area.

In the Doncaster earthquake of 1905, it was felt over most of the isoseismal 4 (area, 10,700 square miles; disturbed area, 17,000 square miles). In the earthquakes listed in Table II, it was noticed over a

part only of the disturbed area. In these cases, the area over which the twin shock was felt ranges from one-third to four-fifths of the area of the isoseismal 4.

TABLE II

Earthquake	Twin-Shock Area in Square Miles	Area of Isoseismal 4 in Square Miles	Disturbed Area
Hereford, 1926	41,500	90,000	90,000
Swansea, 1906	13,500	37,800	66,700
Derby, 1904	8,000	10,120	25,000
Carlisle, 1901	1,130	3,700	3,700

When the places at which the twin shock is observed are plotted on a map, they are seen to lie almost uniformly over the greater part of the disturbed area.[1] Though observers often fail to notice the twin shock, it was recorded by 68 per cent of them in the Derby earthquake of 1903; by 32 per cent in the Doncaster earthquake of 1905; 23 per cent in the Hereford earthquake of 1926; 21 per cent in that of 1896; about 20 per cent in the Derby earthquake of 1904; and 16 per cent in the Swansea earthquake of 1906.

Nor do the percentages vary much in successive isoseismal zones. For instance, in the Hereford earthquake of 1896 the percentage was 20 within the isoseismal 8; and 23, 18, 22, and 28 in successive isoseismal zones. In the Hereford earthquake of 1926, it was 21 within the isoseismal 6; and 24 and 18 in the surrounding zones. In the Doncaster earthquake of 1905, it was 31 within the isoseismal 7; and 38 and 32 in the surrounding zones.

Coalescence of the two parts of the shock.—In some twin earthquakes, though not in all, the twin shock area is crossed by a narrow band within which the two parts of the shock coalesce and form a continuous series of vibrations. The boundaries of this band—the synkinetic band—are shown in Figures 2 and 3, the band being straight and about 5 miles wide in the Derby earthquake of 1903 and curved and from 2 to 4 miles wide in the Stafford earthquake of 1916. In three other earthquakes, the axial line only of the band can be drawn; and, in each case, the line is curved. These earth-

[1] For the Hereford earthquake of 1896, see Fig. 11, p. 231, of my report on that earthquake.

quakes are the Hereford earthquakes of 1896 and 1926 and the Derby earthquake of 1904. For the Derby earthquake of 1906, the observations are too few to determine the existence of a synkinetic band. There is, however, no trace of such a band in the Pembroke earthquake of 1892, the Leicester and Carmarthen earthquakes of 1893, the Carlisle earthquake of 1901, the Doncaster earthquake of 1905, or the Swansea earthquake of 1906.

Mean duration of the interval between the two parts of the shock.—Usually, the duration of the interval lies between 2 and 3 seconds. For the whole disturbed area, the average interval is 2.1 seconds in the Derby earthquake of 1904 and the Swansea earthquake of 1906; 2.2 seconds in the Stafford earthquake of 1916; 2.3 seconds in the Carmarthen earthquake of 1893; 2.4 seconds in the Hereford earthquake of 1896; 2.5 seconds in the Leicester earthquake of 1893; 2.8 seconds in the Hereford earthquake of 1926; 3.0 seconds in the Pembroke earthquake of 1892, the Carlisle earthquake of 1901, and the Derby earthquake of 1903; and 3.5 seconds in the Doncaster earthquake of 1905.

Not only is the interval roughly constant from earthquake to earthquake, but it varies but slightly throughout the disturbed area of any one earthquake. The slight increase with distance that is

TABLE III

Earthquake	A	B	C	D	E
Hereford, 1896	2.4	2.4	2.2	3.0	3.2
Swansea, 1906	2.2	2.1	2.0	2.2	.....
Stafford, 1916	2.0	2.1	2.3	2.3	1.1
Hereford, 1926	.....	2.2	2.9	3.0	.....

sometimes apparent is, I imagine, no more than might be expected from the extinction of the weak terminal vibrations of both parts of the shock. In Table III, the mean durations of the interval are given in seconds for the innermost isoseismal (A) and for the zones between successive pairs of isoseismals.

Relative intensity of the two parts of the shock.—Taking first the 5 earthquakes in which a synkinetic band is traceable, in the Hereford earthquakes of 1896 and 1926, we have isoseismal lines with their longer axes approximately northwest and the synkinetic bands con-

cave toward the southeast. In the former earthquake, the first part of the shock was the stronger on the northwest side of the synkinetic band, and the second part on the southeast side. In the earthquake of 1926, the second part was the stronger on the northwest side, and the first part on the southeast side. In the Derby earthquake of 1903, the two parts of the shock were nearly equal in strength all over the disturbed area, though, as in the Derby earthquake of 1904, the first part was slightly the stronger on the southwest side of the synkinetic band. In the Stafford earthquake of 1916, the difference in strength was slight, but the first part was the stronger on the east side of the synkinetic band and the second part on the west side.

In earthquakes without a synkinetic band, the distribution of relative intensity is different. In the Doncaster earthquake of 1905, as we have seen, the first part was the stronger all over the disturbed area, except within a small and nearly circular area including the eastern portion of the isoseismal 7. In the Swansea earthquake of 1906, the second part was much the stronger in all parts of the disturbed area, except within a circular area near the eastern end of the isoseismal 8. In the Carmarthen earthquake of 1893, the second part was the stronger near the western end of the isoseismal 6; the first part elsewhere.

Relative positions of the isoseismal lines.—The twin earthquakes described above show the various relative positions of the isoseismal lines. In the Doncaster earthquake of 1905, the isoseismal 7 consists of two detached portions. In the Leicester earthquake of 1893, the innermost isoseismal is excentric with regard to the others; and this is also the case in the Carlisle earthquake of 1901, the Derby earthquake of 1904, and the Swansea earthquake of 1906. In the other 7 twin earthquakes, the relations between the innermost isoseismals are normal.

TWIN EARTHQUAKES CONNECTED WITH TWO DETACHED FOCI

The two series of vibrations which form a twin earthquake might, it is conceivable, be due to several causes, such as (1) a single impulse repeated by reflection or refraction at the bounding surfaces of different strata; (2) the separation of condensational and distortional waves as they advance outward from the origin; (3) a repetition of

the impulse within the same or an overlapping focus; or (4) the occurrence of simultaneous, or nearly simultaneous, impulses in two detached foci.

1. The first explanation is excluded by the wide area over which the twin shock is felt, the general uniformity of its observation throughout the disturbed area, the existence in some earthquakes of a synkinetic band, and the definite laws of distribution of the relative intensity of the parts.

2. To the second explanation, the existence of the synkinetic band is equally opposed, as well as the distribution in the relative intensity of the two parts. Moreover, in the central district, the twin shock would be confused or absent and the interval between the two parts should increase in proportion to the distance from the center. Further, if the second series consisted of distortional vibrations, every earthquake would be a twin earthquake.

3. Nor can twin earthquakes be due to a repetition of the impulse within the same or an overlapping focus; for, if this were the case, the order of relative intensity would not vary according to a definite law throughout the disturbed area, there would never be a synkinetic band, and the mean duration of the interval would not be confined within the narrow limits of 2 or 3 seconds.

4. On the other hand, all the known phenomena of twin earthquakes are readily explained by the supposition that they are due to nearly simultaneous impulses in two detached, or almost detached, foci. The twin shock would be perceptible as far as the weaker of the two parts can be felt, and over the whole disturbed area if the impulses differed but little in strength. If the impulses were precisely simultaneous, the vibrations from the two foci would coalesce along a straight synkinetic band, as in the Derby earthquake of 1903. If the second impulse were to follow the first after an interval less than that required for the waves from the first to traverse the interfocal distance, there would still be a synkinetic band; but it would be hyperbolic in form and its concavity would face the focus last in action, for the waves from the second focus would have a shorter distance to travel than those from the first before the two coalesced. As we have seen, the synkinetic band was hyperbolic in the Hereford earthquakes of 1896 and 1926, the Derby earthquake

of 1904, and the Stafford earthquake of 1916; and, in each of these earthquakes, the synkinetic band separated the areas of opposite relative intensity. Again, if the second impulse were to follow the first after an interval greater than that mentioned above, there could be no synkinetic band; and either the first or the second part would be the stronger all over the disturbed area, except that in the immediate neighborhood of the focus with the weaker impulse there might be a small area in which the vibrations from that focus would be felt the more strongly—as in the Carmarthen earthquake of 1893, the Doncaster earthquake of 1905, and the Swansea earthquake of 1906. Lastly, the relative position of the isoseismal lines in some twin earthquakes points to a double focus. In the Doncaster earthquake of 1905, the two epicenters no doubt coincide with the centers of the two portions of the isoseismal 7. In the Leicester earthquake of 1893, the Carlisle earthquake of 1901, the Derby earthquake of 1904, and the Swansea earthquake of 1906 the center of the excentric innermost isoseismal marks the epicenter corresponding to the stronger impulse. In the remaining earthquakes, the impulses were so nearly equal in strength and the foci so close together that the innermost isoseismal surrounds both epicenters.

This reasoning is strongly supported by the form of the isacoustic lines. Two such lines are shown in the map of the Derby earthquake of 1903 (Fig. 2), and they are strongly distorted in the direction of the synkinetic band. Isacoustic lines, similarly distorted, have been drawn for the Hereford earthquakes of 1896 and 1926 and the Derby earthquake of 1904. For the Doncaster earthquake of 1905 and the Swansea earthquake of 1906 the lines were not actually drawn owing to the lack of sufficient observations; but, from the distribution of the audibility-percentages in certain areas, it is clear that there would be no distortion of the isacoustic lines in the directions at right angles to the longer axes of the isoseismals. When such distortion along the synkinetic band does occur, it is no doubt due to the coalescence of the sound-vibrations from the two foci of the earthquake.

As a rule, the position of the twin epicenters can be determined within about a mile. For instance, in the Derby earthquake of 1903, one epicenter is 1½ miles east of Ashbourne, the other 3 miles west of Wirksworth. In the Stafford earthquake of 1916, the epicenters

lie about 2 miles northeast of Stafford and 1½ miles northwest of Eccleshall. In the Doncaster earthquake of 1905, they are ½ mile north of Bawtry and 4 miles east of Crowle.

CLASSIFICATION OF TWIN EARTHQUAKES

It will be seen that the 12 twin earthquakes considered in this paper differ in several respects so distinctly that they may be arranged in two classes. The following are the chief points of difference:

1. In the first place, we note the presence or absence of a synkinetic band, coupled with the presence or absence of distortion of the isacoustic lines. The band exists in at least 5 earthquakes, namely, the Hereford earthquakes of 1896 and 1926, the Derby earthquakes of 1903 and 1904, and the Stafford earthquake of 1916. For the Derby earthquake of 1906 the observations are insufficient to determine its existence. In the remaining earthquakes—the Pembroke earthquake of 1892, the Leicester and Carmarthen earthquakes of 1893, the Carlisle earthquake of 1901, the Doncaster earthquake of 1905, and the Swansea earthquake of 1906—there is no trace whatever of a synkinetic band.

2. There can be little doubt that this difference is connected with the difference in the distance between the two epicenters. Approximately, the distance is 9 miles in the Hereford earthquakes of 1896 and 1926, and 8 or 9 miles in the Derby earthquakes of 1903 and 1904 and the Stafford earthquake of 1916. On the other hand, it is 17 miles in the Leicester earthquake of 1893 and the Doncaster earthquake of 1905, 22½ miles in the Swansea earthquake of 1906, and 23 miles in the Carlisle earthquake of 1901. In the twin earthquakes of the midland counties, the distance between the epicenters in one class (about 8½ miles) is just half that between the epicenters in the other (17 miles).

3. In earthquakes of the first class, the interval between the two parts of the shock is one of absolute rest and quiet. It is only within, and near the margins of, the synkinetic band that the shock consists of two maxima connected by weaker tremulous motion. In those of the second class, the shock in the central region nearly always consists of two maxima joined by tremulous motion; though, at some

distance from the epicenters, the tremulous motion becomes imperceptible, and the shock consists of two distinct parts separated by an interval of rest and quiet.

4. Twin earthquakes of the first class are usually followed after a short interval by a slight shock in the interfocal region. The interval between the principal earthquake and the interfocal shock varies much, being about 10 minutes for the Hereford earthquake of 1896, 8 hours for the Derby earthquake of 1904, 15 hours for the Stafford earthquake of 1916, and 40 days for the Derby earthquake of 1903.

5. In twin earthquakes of the first class, foreshocks or aftershocks sometimes occur in one focus only. In those of the second class, an earthquake may occur in a single focus quite apart from a twin earthquake. For instance, the Doncaster earthquake of April 23, 1905, was preceded on April 13, 1902, by a slight shock (intensity 4) in the northeastern focus (Fig. 4). The Swansea earthquake of June 27, 1906, was followed on July 3, 1907, by a slight shock (intensity 4) in the western focus only. Lastly, the Leicester earthquake of August 4, 1893, was followed on June 21, 1904, by a slight shock (intensity 3) in the northwestern focus near Loughborough; and this again, 2 hours later, by a stronger shock (intensity 5) in the other focus 17 miles to the southeast (Fig. 5).

Yet, notwithstanding the differences noticed above, there are important and close points of resemblance: (1) in their nature, especially at some distance from the origin; (2) in their large disturbed areas (indicative of great depth of origin), which, for earthquakes of intensities 8 and 7, are on an average 44,500 square miles for twin earthquakes of the first class and 32,300 square miles for those of the second class, as compared with 11,200 square miles for simple earthquakes; and (3) in the number of accessory shocks, the average being 4.0 for twin earthquakes of the first class, 3.5 for those of the second class, as compared with 12.0 for simple earthquakes.

ORIGIN OF TWIN EARTHQUAKES

It is obvious that the origins of twin earthquakes of the two classes here described cannot be identical. Taking, first, the earthquakes in which a synkinetic band exists, we have to account for the simultaneous, or almost simultaneous, occurrence of impulses in

two detached foci. The second impulse cannot be a consequence of the first, for it takes place before the waves from the focus first in action have time to reach the other. Yet they cannot be independent, for the chances against two movements occurring within 2 or 3 seconds of one another, and nearly always within 2 or 3 seconds, are practically infinite. For, in the Staffordshire foci, no other earthquake is known to have occurred during the last thousand years. In the Derbyshire foci, the chances against independent movements are higher still, for in them practically simultaneous movements have recently occurred three times within little more than three years

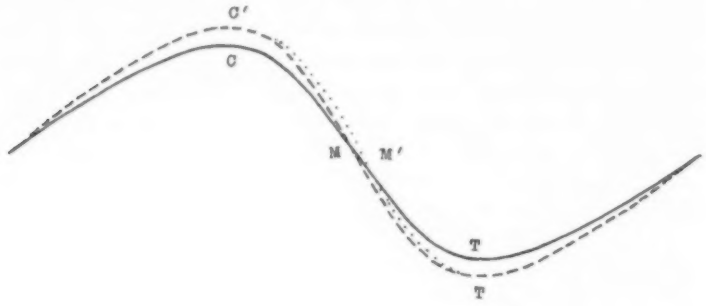

FIG. 6.—Origin of twin earthquakes

(March, 1903–August, 1906). Yet, for a thousand years we have otherwise no recorded movement within or near the same twin foci.

There is, however, one single movement that can produce such nearly simultaneous displacements in detached areas, and that is a movement of rotation, such a rotation as would occur, for example, in the growth of a deep-seated crust fold. In Figure 6, the curve *CMT* is supposed to represent a section of a fold by a transverse fault, the crest *C* and the trough *T* being separated by a distance of 8 or 9 miles. If a small step were to take place in the growth of the fold, that is, if the crest and trough were to become slightly more pronounced, simultaneous movements of the fold—the crest *C* to *C'* and the trough *T* to *T'*—would occur; and these movements would be accompanied by a rotation of the median limb *M*, the central portion of which would undergo no displacement. Thus, we should

have two foci, CC' and TT', entirely separated by the interfocal region about M.

Such a step in the growth of the fold would leave the median limb subjected at both ends to increased stresses. As the initial impulses are, as a rule, unequal, the stresses should afterward be relieved by a slip in the interfocal region, a translation of the median limb from M to M', as if the growth of the arches were followed by a smaller bodily advance of the fold. When the initial impulses in the two foci are nearly equal in strength, as in the Derby earthquake of 1903, the afterslip may be delayed for 40 days. When they differ in strength, the interval may be brief—10 minutes in the Hereford earthquake of 1896, 8 hours in the Derby earthquake of 1904, and 15 hours in the Stafford earthquake of 1916.

In twin earthquakes of the second class, the distance between the epicenters is roughly double that in earthquakes of the first class; and it may therefore be supposed that the foci occupy similar positions, say on the crests of two successive folds. As the shock near the center is unbroken, the movement in the first focus seems to be continued slightly across the interfocal region to the second focus. And, further, as there is no synkinetic band, it follows that the vibrations from the first focus reach the second focus before the latter comes into action. Thus, while motion may be just on the point of taking place in both foci, the movement in the second focus is precipitated by that in the first.

Why twin earthquakes should be sometimes of one class and sometimes of the other is by no means clear. Is it possible that, under certain conditions, the trough of a fold may be unable to increase in depth, and that the movement is thus continued on to the next fold? There does seem to be a progressive movement as a whole from fold to fold, for, in the midland counties of England, we have first the twin earthquake of Northampton in 1750; then the Leicester earthquake in 1893, followed by the Derby earthquakes in 1903, 1904, and 1906; and finally by the Stafford earthquake in 1916.

In this concluding section, some other properties of twin earthquakes may be referred to more appropriately than in the earlier sections.

Stability of twin earthquake foci.—Simple earthquakes are marked

by a continual migration of the focus, an oscillation from side to side along the fault surface, with a concluding tendency in Great Britain either to a westerly migration or a return to the central region. In twin earthquakes, on the other hand, there seems to be little, if any, perceptible migration of the foci. The Hereford earthquakes of 1863, 1896, and 1926, and possibly also that of 1868, originated in the same twin foci, one near Hereford, the other about 9 miles to the southeast, near Ross. The Derby earthquakes of 1903, 1904, and 1906 originated in the same twin foci, one near Ashbourne, the other 8 or 9 miles to the north-northeast near Wirksworth; and so also the Leicester earthquakes of 1893 and 1904 in foci, one near Loughborough, the other 17 miles to the southeast near Burton Overy.

Order of relative intensity of the two impulses.—There seems to be no discernible law governing the order of relative intensity of the two impulses, though in 7 out of 11 twin earthquakes the earlier impulse was the stronger, namely, the Leicester and Carmarthen earthquakes of 1893, the Hereford earthquake of 1896, the Carlisle earthquake of 1901, the Doncaster earthquake of 1905, the Derby earthquake of 1906, and the Stafford earthquake of 1916. In 4 earthquakes—the Pembroke earthquake of 1892, the Derby earthquake of 1904, the Swansea earthquake of 1906, and the Hereford earthquake of 1926—the second impulse was the stronger.

The same diversity prevails in the same twin foci. In each of the Hereford earthquakes of 1863, 1896, and 1926, the Hereford focus was first in action. In 1863 and 1926, the impulse in the Ross focus was the stronger; in 1896, that in the Hereford focus. In each of the Derby earthquakes of 1903, 1904, and 1906, the impulse in the Ashbourne focus was the stronger, though very slightly so in 1903. In this year, the impulses were simultaneous; in 1904, the Wirksworth focus was first in action; and in 1906, the Ashbourne focus.

Deep-seated foci of twin earthquakes.—In the whole of Scotland and in northern and central Wales, no twin earthquake is known to have occurred. The great majority of earthquakes in Scotland can be referred to known faults. They must therefore be of superficial origin, and, in successive aftershocks of strong simple earthquakes—as, for instance, the Inverness earthquakes of 1890 and 1901—it is possible to trace the continual approach of the foci toward the

surface. Some of the simple earthquakes of England and Wales may also be so associated, such as the Bolton earthquake of 1889, the Carnarvon and Bala earthquakes of 1903, the Malvern earthquake of 1907, and the Barnstaple earthquakes of 1894 and 1902. But, with one exception—the Leicester earthquake of 1893—it is impossible to refer any twin earthquake to a known fault.

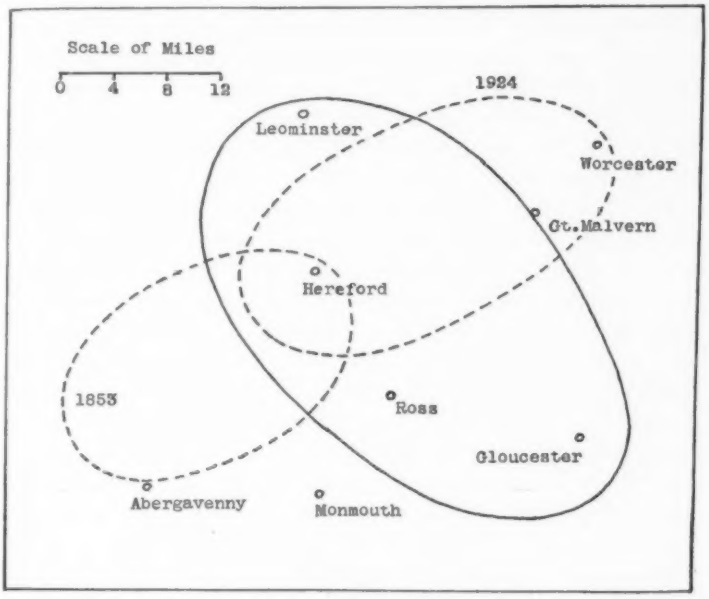

FIG. 7.—Hereford earthquakes of 1853 and 1924

For earthquakes of the same intensity, as we have seen, twin earthquakes disturb a much larger area than simple earthquakes, an area about three times as great for those of intensities 8 and 7 and between four and five times as great for those of intensity 5. For accessory shocks, the inequality is even more marked, the ratio being more than 8:1; and we have such remarkable areas as over 6,000 square miles for two foreshocks (intensity about 4) of the Hereford earthquake of 1896 and nearly 30,000 square miles for an aftershock (intensity about 5) of the Pembroke earthquake of 1892.

Though definite figures cannot be given, it may, I think, be assumed that the focal depths of most twin earthquakes are of the order of 5 or 10 miles.

Effects of twin earthquake movements on the surface crust.—Yet, deep-seated though they are, twin earthquake movements affect the stability of the surface crust. Some time afterward—it may be days or it may be years afterward—twin earthquakes are followed by movements along a surface fault crossing the central district in a direction nearly at right angles to the axes of the isoseismal lines of the twin earthquake. For instance, the Pembroke earthquake of 1892, which originated along a deep north-and-south fault, was succeeded after less than 4 hours by an earthquake along an east-and-west fault traversing the central district, and this again by another 4 days later. The Carlisle earthquake of 1901, due to movements along a deep north-and-south fault, was followed 10 years later (on May 16, 1911) by a movement in a nearly east-and-west fault passing close to the southern epicenter of the twin earthquake near Grasmere. The most interesting example is, perhaps, that of the Hereford earthquakes. The continuous line in Figure 7 represents the isoseismal 8 of the earthquake of 1896. The dotted lines are the boundaries of the disturbed areas of two slight shocks—on March 27, 1853, and January 26, 1924—the axes of which pass through the Hereford epicenter of 1896. The former may be the successor of some long-past movement; the latter, of the earthquake of 1896. In the first two examples, the axes of the superficial earthquakes were crossed centrally by the axes of the preceding twin earthquakes (of the second class); in the Hereford earthquakes of 1853 and 1924, the axes lie on either side of the axis of the related twin earthquake (of the first class).

The superficial origin of these movements is indicated by the association of the Pembroke earthquake with a known fault; by the small disturbed area (112 square miles) of the second of these earthquakes as compared with that of another aftershock (about 4,800 square miles) connected with the main fault and of approximately the same intensity on the surface; and by the elongated, but small disturbed areas of all of them and by the closeness of the isoseismal lines of the Grasmere earthquake of 1911.

THE ORIGIN AND OCCURRENCE IN ROCKBRIDGE COUNTY, VIRGINIA, OF SO-CALLED "BENTONITE"

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ABSTRACT

Certain strata found south of Lexington, Rockbridge County, Virginia, in the upper part of the Liberty Hall formation of late Middle Ordovician age have been interpreted as "bentonite" and of volcanic origin. Careful study of the material indicates an almost total absence of the properties normally attributed to bentonite. The material consists of minerals related to the clay group and exhibits slightly schistose structure both megascopically and microscopically rather than the structure inherited from a tuff. It would seem to be classified conservatively and appropriately as a high-alumina shale with slightly schistose structure produced by dynamic metamorphism. Nor is the assumption of volcanic origin strengthened by the postulation that the feeding vent was located east of the Blue Ridge in Nelson County, Virginia, the rocks of which are clearly plutonic in origin and probably pre-Cambrian in age.

INTRODUCTION

During the past fifteen years a number of descriptions of supposed volcanic ash beds found in the Ordovician of eastern United States have appeared. Taber seems to have been the first to direct attention to these deposits in his description of a tuff in the Arvonia (Martinsburg) slates of Buckingham County, Virginia, east of the Blue Ridge.[1] Many years ago Ulrich described a soft clay layer in the Ordovician at High Bridge, Kentucky, which since has been interpreted as bentonite.[2] Nelson has repeatedly described deposits, designated as bentonite, found in Tennessee, Alabama, and Virginia[3] and has added to their interest by attempting to locate with positive-

[1] Stephen Taber, "Geology of the Gold Belt in the James River Basin, Virginia," *Bull. 7, Va. Geol. Survey* (1913), p. 43.

[2] E. O. Ulrich, "Correlation of the Lower Silurian Horizons of Tennessee and the Ohio and Mississippi Valleys with Those of New York and Canada," *Amer. Geologist*, Vol. I (1888), p. 107.

[3] Wilbur Nelson, "Notes on a Volcanic Ash Bed in Middle Tennessee," *Bull. 25, Tenn. Geol. Survey* (1921), pp. 46–48; "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), pp. 605–16; *Bull. 27, Tenn. Geol. Survey* (1923), p. 23; "Appalachian Bauxite Deposits," *Bull. Geol. Soc. of Amer.*, Vol. XXXIV (1923), pp. 525–40; *Bull. 35, Tenn. Geol. Survey* (1925), p. 31; "Volcanic Ash Deposit in the Ordovician of Virginia," *Bull. Geol. Soc. of Amer.*, Vol. XXXVII (1926), pp. 149–50.

ness their source.[1] In Alabama, according to Butts, such deposits have been found 1½ miles southeast of Ragland, St. Clair County, on Red Mountain near Attalla, at Birmingham, and 3 miles southeast of Bessemer.[2] In northwest Georgia "there are certain clay-like materials associated with the Chickamauga formation which may possibly be correlated with Nelson's so-called bentonite."[3] Deposits interpreted as bentonite and supposed to be of volcanic origin have been found at many places in eastern and central Tennessee.

Nelson has described a bed which he called "bentonite" south of Lexington, Rockbridge County, Virginia, on the Dixie Highway leading to Natural Bridge.[4] During the summer of 1926 numerous other so-called "bentonite" beds were reported by the Virginia Geological Survey from various places west of the Blue Ridge in Virginia.[5] Bucher reports the occurrence of bentonite in a well in Perry County, Ohio.[6] Sardeson has described occurrences of a clay lay which he interpreted as bentonite in northwestern Wisconsin a Ellsworth and in the adjacent part of Minnesota. He correlates 'h' bed with similar material found in the Decorah shale at St. Pau and with the Appalachian bentonite deposits.[7] Recently Stose and Jonas have noted the occurrence of an amygdaloid basalt flow in association with spotted green and purple shales or slates of probable volcanic origin found near Jonestown, Pennsylvania, and have suggested the possible locus of a feeding vent.[8]

[1] Wilbur Nelson, "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), pp. 605–16; "Volcanic Ash Deposit in the Ordovician of Virginia," *Bull. Geol. Soc. of Amer.*, Vol. XXXVII (1926), pp. 149–50.

[2] Charles Butts, *et al.*, "The Geology of Alabama," *Special Report No. 14, Alabama Geol. Survey* (1926), pp. 113–14, 131.

[3] Personal communication from S. W. McCallie, state geologist of Georgia.

[4] Wilbur Nelson, "Volcanic Ash Deposit in the Ordovician of Virginia," *Bull. Geol. Soc. of Amer.*, Vol. XXXVII (1926), pp. 149–50.

[5] Personal communication from Dr. J. S. Grasty, Charlottesville, Virginia.

[6] Wilbur Nelson, "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), pp. 606.

[7] F. W. Sardeson, "Volcanic Ash in Ordovician Rocks in Minnesota," *Pan-American Geologist*, Vol. XLII (1924), pp. 45–52; "Beloit Formation and Bentonite," *Pan-American Geologist*, Vol. XLV (1926), pp. 383–92; Vol. XLVI, pp. 11–24.

[8] George Stose and I. Jonas, "Ordovician Shale and Lava in Southeastern Pennsylvania," *Bull. Geol. Soc. of Amer.*, Vol. XXXVIII (1927); also personal communication.

As interpreted by the various authors, the deposits range in age from early Ordovician (St. Peter sandstone and Little Oak limestone) to the middle Cincinnatian (Buckingham tuff). The span of geologic time represented in this wide divergence in geologic horizon has a significant bearing on the origin of the beds, particularly as periods of volcanic activity affecting individual vents are of limited duration geologically.

Since vulcanism is one of the dominant processes of geologic history and since igneous rocks are consequently widespread, descriptions of isolated occurrences necessarily attract little attention. The classification, however, of these Ordovician deposits as "bentonite" is of especial interest as bentonite is essentially confined to rocks of late Mesozoic and Tertiary age. If this classification is correct, these beds should be not only of economic importance but also of scientific interest in the bearing their classification would have upon the classification of many Paleozoic and later beds of volcanic origin and described simply as "tuff," "altered tuff," "halleflintas," "volcanic slates," etc. The establishment of their origin as volcanic would also be of considerable scientific significance because of the general absence of volcanism in eastern United States during the Ordovician.

ORIGIN OF THE "BENTONITES"

Professor Nelson, who has most frequently described the "bentonites," early erected a supposititious volcano of Lowville (early Black River) age between Elliott and Fayette counties, Kentucky, thus affording the deposits of supposed volcanic origin an appropriate and legitimate parentage.[1] Schuchert has named the volcano "Nelson's Volcano" for its founder.[2] The announcement of the volcano created considerable scientific interest chiefly because of its postulated dimensions. Professor Nelson has estimated that the quantity of ash erupted would comprise a volume of 66 cubic miles, the material being spread over eastern United States reaching places as far away as Arkansas and Wisconsin.[3]

[1] Wilbur Nelson, "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), pp. 605–16.

[2] *Historical Geology* (1924), p. 239.

[3] Wilbur Nelson, "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), pp. 605–16.

The subsequent discovery of supposed ash deposits interpreted as bentonite has led to repeated relocations of the great volcano. In December, 1925, the Rockbridge "bentonite" was described and the volcano relocated "under the Cumberland mountains near the point where Kentucky, West Virginia, and Virginia meet."[1] In May, 1926, before the Virginia Academy of Science, meeting in Charlottesville, Professor Nelson, state geologist of Virginia, again described the Rockbridge material and located the so-called "Nelson's Volcano" just south of the University of Virginia on the Nelsonite comagmatic complex of the Nelson County area.[2] The location of an Ordovician volcano in a region where there had previously been no intimation that such a feature was possible; and where in fact it had previously been demonstrated by Watson and Taber to be highly improbable, naturally aroused the interest of the writer and others who have long studied the geology of Virginia.

The location of the supposititious Middle Ordovician volcano in the rutile area of central Virginia was peculiarly unfortunate. The rocks of that region had been described by Watson and Taber fifteen years ago in a monograph that is a classic in American geology.[3] These authors have demonstrated that the rocks of the region are plutonic in origin and resulted from magmatic differentiation in pre-Cambrian times. Professor Taber informs the writer that "the igneous rocks in the Nelson County area are clearly plutonic in origin and not volcanic."[4]

These authors state that in this comagmatic region:

> Five principal rock groups have been recognized and described, which, named in their probable order of differentiation, including the surrounding rocks, are: (1) Biotite-quartz monzonite-gneiss with schists, (2) syenite, (3) gabbro, (4) nelsonite, and (5) diabase.[5]

[1] Wilbur Nelson, "Volcanic Ash Deposit in the Ordovician of Virginia," *Bull. Geol. Soc. of Amer.*, Vol. XXXVII (1926), pp. 149–50.

[2] Personal communications from Dr. J. S. Grasty; Dr. Harry Campbell; Dr. W. J. Humphries; Dr. A. A. Pegau; Mr. C. E. Bass, of the University of Virginia; and Mr. Dewey Fleming, of the editorial staff of the *Baltimore Sun*.

[3] T. L. Watson and Stephen Taber, "Geology of the Titanium and Apatite Deposits of Virginia," *Bull. 3-A, Va. Geol. Survey* (1913).

[4] Personal communication from Dr. Stephen Taber.

[5] Watson and Taber, *op. cit.*, p. 230.

The rocks of the region are all intrusive and holocrystalline in texture. Most of them are even-granular, ranging from very coarse-grained in the syenite to very fine-grained in some of the nelsonite and diabase dikes. Porphyritic texture is observed in some of the nelsonite and gabbro-nelsonite dikes. The bulk of the rocks, including quartz monzonite-gneiss, syenite (feldspathic and hornblendic border facies), and a part of the gabbros and nelsonites are deep-seated (plutonic) masses, traversed by dikes of gabbro, diabase, and nelsonite. These are genetically related, and with the exception of diabase nearly all gradations in composition are indicated between the extremes of the different types.

The indications from field study are that the syenite and deep-seated gabbro usually grade into each other, sharp contacts being but seldom observed. Likewise between the gabbro and nelsonite occur nearly all gradations in mineral composition, from typical hypersthene gabbro (norite) containing subordinate apatite and ilmenite to typical nelsonite containing little or none of the silicate minerals.

The rocks exhibit pronounced though unequal effects of dynamic metamorphism, both in hand specimens and in thin sections under the microscope. The most pronounced effect of metamorphism visible in hand specimens of most of the rocks is the development of complete or partial banded or gneissic structure. Microscopically metamorphism is manifested chiefly in mashing—granulation, fractures, and optical disturbance of certain essential minerals—in recrystallization, and in the complete or partial change of pyroxene (hypersthene) to secondary hornblende (actinolite).[1]

The quartz-biotite monzonite-gneiss is the dominant type of the southeast slope of the Blue Ridge and southeastward along its base, in middle western Virginia. Just beyond the southeast limits of the mapped area is an extensive belt of Lower Cambrian rocks, the lowest members of which have been formed from detritus derived from the quartz-biotite monzonite-gneiss. The evidence seems conclusive, therefore, that the gneiss at least, which is the country rock of the region mapped, is pre-Cambrian. Until more conclusive evidence can be found, which must be sought beyond the limits of the area mapped, the remaining rock types, with probably the exception of diabase, must likewise be regarded as pre-Cambrian."[2]

The region is mapped as pre-Cambrian in the report and also on the widely distributed Virginia Geological Map.

The writer has made a special study of the Rockbridge "bentonite,"[3] the origin of which was attributed to an eruption of the volcano located in the rutile area of Nelson-Amherst counties.

[1] Watson and Taber, *op. cit.*, pp. 202–3.

[2] *Ibid.*, p. 231.

[3] Wilbur Nelson, "Volcanic Ash Deposit in the Ordovician of Virginia," *Bull. Geol. Soc. of Amer.*, Vol. XXXVII (1926), pp. 149–50.

AGE OF THE ROCKBRIDGE "BENTONITE"

A part of a day during the summer of 1925 was spent in the field with Dr. J. S. Grasty measuring the section where the Rockbridge "bentonite" was reported to be present, but the deposit was not found. Later it was pointed out to Professor Harry Campbell, who had been unsuccessful after prolonged and persistent search in locating the elusive "bentonite."[1]

Professor Campbell assigns the horizon of the Rockbridge "bentonite" to the Liberty Hall formation of Upper Black River age. The formation is about 1,000 feet thick in Rockbridge County and consists chiefly of dark argillaceous strata.

> The Liberty Hall limestone is usually a succession of rather evenly banded beds of fine-grained, dark blue limestone, and darker, more argillaceous limestone which weathers shaly. As we ascend into the formation calcareous shale predominates and limestone beds are less frequent. In this region the formation has been much fractured and folded, and sometimes appears massive with innumerable veins of infiltration of calcite filling the crevices. Again it appears shaly after long exposure to the weather.[2]

The following stratigraphic section (2.7 to about 5.5 miles from Lexington Court House) reveals the rock succession. This section was subsequently checked by Dr. Campbell.

STRATIGRAPHIC SECTION SOUTH OF LEXINGTON, VIRGINIA

Liberty Hall Limestone
> Thin bedded, contorted and shaly limestone with gray siliceous shales and thin bands of limestone (carries the "bentonite" bed).

Athens shale?
> Yellowish-gray shale.

(Fault)

Murat limestone (5.4 miles from Lexington Court House)
> Gray contorted limestone.

(Fault)

Natural Bridge limestone
> Gray and bluish-gray, massive, laminated limestone.

Murat limestone (5.6 miles from Lexington Court House)
> Contorted, bluish-limestone, knotty. Folded or faulted down into Natural Bridge limestone.

Natural Bridge limestone
> Gray and bluish-gray, massive, shaly and laminated limestone.

[1] Personal communication from Dr. Harry Campbell.

[2] R. S. Bassler, "The Cement Resources of Virginia West of the Blue Ridge," *Bull. 2-A, Va. Geol. Survey*, p. 109.

The layer of "bentonite" is in the upper shaly part of the Liberty Hall formation at a point between 4.8 and 4.9 miles from Lexington Court House on the southwest side of Elliott's Hill. It is some distance from the top of the hill and about 200 yards from a sharp turn in the road close to a small concrete bridge.[1]

The brownish-yellow deposit is 10 to 12 feet thick with strike N. 27° E. and dip 75° SE. It is "scarcely distinguishable from the shale on either side," according to Professor Campbell. A thick bed of chert lies against the "bentonite." The strata in the general vicinity have been much deformed both by folding and by faulting, and possess high dips.

Twenty-five feet south of the "bentonite" bed there is a bed of similar yellowish decayed shale 2 to 3 feet thick; and 100 feet beyond the "bentonite" there is another similar shale bed 2 feet thick with a chert bed 6 inches thick in contact with it on the southeast. "At a point 5.2 miles from the Court House and adjoining a two-foot cherty bed, there are about 8 feet of shaly material of a yellowish to greenish tone."[2] This indicates how shaly the Liberty Hall formation is in this vicinity and the similarity of many beds in the section. All beds possess high dips, some being vertical.

DESCRIPTIONS OF THE "BENTONITES"

Immediately after the location of the Rockbridge material interpreted as bentonite had been pointed out to Dr. Campbell by Professor Nelson, the former sent the writer a large sample from the deposit. Later Dr. J. S. Grasty mailed the writer a sample representing an 8-foot cross-cut of the bed. The writer's study has been based upon these collections.

Microscopic examination of the Rockbridge material reveals clearly that it is a mixture and consists of minerals related to the clay group, though one mineral predominates. This mineral exhibits lamellar orientation, excellent cleavage in one direction, and high refractive indices. Biotite fragments are common. Quartz grains are rare, and feldspar grains are relatively rare. Since the sediments of the Valley of Virginia have been derived from a crystalline land mass

[1] Personal communication from Dr. Harry Campbell.

[2] Personal communication from Dr. J. S. Grasty.

to the east, such accessory minerals may be expected in essentially all microscopic sections. Megascopically the material exhibits slightly schistose structure, and the microscopic sections indicate to a considerable extent the structural nature of the material to be that of altered shale rather than the structure inherited from volcanic tuff.

In an effort to compare the Rockbridge "bentonite" with other Ordovician bentonites, a survey of all the literature has been made. While the references are numerous, the information gleaned was meager. Little can be gained in the perusal of the scattered notes except the name of the formation and the thickness of the bed. The only paper that deals with the deposits at any length has much of its space allotted to the erection of the volcano and the account of its supposed eruptions.[1]

Microscopic work on samples from some of these ancient deposits has been done by Larsen, although he has published nothing on them. He has interpreted the essential mineral as leverrierite.[2] Chemical analyses of the bentonites are few and apparently represent previously impurified material. They show considerable diversity in the proportions of the chemical constituents. The best analysis so far published is that by Ross and Shannon on the High Bridge material, which they have interpreted as a potash-bearing bentonite.[3]

In the brief descriptions of the various occurrences little attention has been given to the subjacent and superjacent contacts, although these should be interesting for the light they might shed on the precise origin of the beds and the environmental conditions obtaining at the time of their formation. Taber has briefly but clearly described the megascopic and microscopic characters of the Buckingham tuff, and the High Bridge material has been carefully studied; but with these exceptions these deposits have received little of the detailed chemical and microscopic work so essential for precise and accurate classification of highly weathered materials.

[1] Wilbur Nelson, "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), pp. 605–16.

[2] Wilbur Nelson, "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), p. 614.

[3] C. S. Ross and E. V. Shannon, "The Minerals of Bentonite and Related Clays and Their Physical Properties," *Jour. Amer. Ceramic Soc.*, Vol. IX, No. 2 (1926), pp. 77–96.

DEFINITION OF BENTONITE

Definitions of bentonite have been proposed by Holmes, Ladoo, United States Geological Survey, United States Bureau of Mines, and others. Most definitions as well as descriptions of bentonite emphasize its unique physical properties, such as its very fine-grained texture, high percentage of water, marked adsorptive powers and hyperplasticity.[1] The following definition, taken from the United States Bureau of Mines and published in the *Engineering and Mining Journal*, is typical: "The name bentonite has been applied to a group or series of claylike materials characterized by an alkaline oxide and alkaline earth content of 5 to 10 percent, fine grain size, high absorptive powers, and usually very strong colloidal properties."[2]

The United States Geological Survey has defined bentonite as "a bedded plastic clay which swells immensely upon wetting."[3] Spence quotes the United States Geological Survey in defining bentonite as "a transported, stratified clay, formed by the alteration of volcanic ash shortly after deposition."[4] The alteration is attributed to the hydration of very fine glass particles that have fallen into shallow bodies of water. As hydration progressed, the particles slowly settled out of suspension following flocculation or coagulation. Microscopic examination of thin sections reveals the typical structure of devitrified glassy tuff and indicates that the material is largely composed of a micaceous secondary clay mineral which has been interpreted by Larsen and others as leverrierite, an orthorhombic mineral with the composition of $2Al_2O_3 \cdot 5SiO_2 \cdot 5H_2O$.

Recently Ross and Shannon have concluded that montmorillonite [$(Mg,Ca)O \cdot Al_2O_3 \cdot 5SiO_2 \cdot nH_2O$ (n equals 5–8)] and beidellite

[1] Arthur Holmes, *The Nomenclature of Petrology* (London: Murby, 1920); R. B. Ladoo, *Non-Metallic Minerals* (New York: McGraw & Hill, 1925); H. Ries, *Clays, etc.*, (New York: Wiley, 1912); G. P. Merrill, *The Non-Metallic Minerals* (New York: Wiley, 1910); *U.S. Geol. Survey, Bull. No. 260* (pp. 559–63), No. *285* (pp. 445–47), No. *624* (p. 342); Hugh S. Spence, "Bentonite," *Can. Dept. of Mines, No. 626* (1924).

[2] "Bentonite, Its Production and Utilization," *Engineering and Mining Journal*, November 19, 1921, p. 819.

[3] "Useful Minerals of the United States," *Bull. 624, U.S. Geol. Survey* (1917), p. 356.

[4] Hugh S. Spence, "Bentonite," *Can. Dept. of Mines; No. 626* (1924).

[$Al_2O_3 \cdot 3SiO_2 \cdot nH_2O$ (n equals about 4, and Al_2O_3 may be partly replaced by Fe_2O_3)] are the essential clay minerals of bentonite. They consider leverrierite to be imperfectly known and its place in this clay group undetermined.[1] These authors propose the following definition for bentonite: "Bentonite is a rock composed essentially of a crystalline clay-like mineral formed by devitrification and the accompanying chemical alteration of a glassy igneous material, usually a tuff or volcanic ash; and it often contains variable proportions of accessory crystal grains that were originally phenocrysts in the volcanic glass."[2] Montmorillonite is regarded as developing from quartz-poor rhyolites, trachytes, latites, and quartz latites upon alteration to bentonite; and phonolite and more basic tuffs alter to beidellite bentonite.

There is an apparent lack of uniformity in the opinions of the various authors regarding the definition of bentonite. While most writers emphasize its physical properties, others stress its origin as the integral part of their definition. To some certain materials with certain properties would be bentonite, to others these materials would not be considered bentonite.

The recent work of Ross and Shannon, however, is a substantial effort to get somewhere in the difficult mineralogy of shales and clays. Unfortunately, present knowledge of the physical and chemical properties and of the occurrence and distribution of montmorillonite and beidellite is by no means as extensive and complete as is desirable for the purpose of precise rock classification. Montmorillonite is known to occur in carbonaceous shales, as at the type locality in France, in Fuller's earth, in altered lithium-bearing pegmatites, etc. Beidellite has been found in altered basic tuffs and in gouge clays of ore veins. Both minerals are probably widely distributed in clays, shales, muds, etc., of diverse origins; hence the presence of these minerals alone cannot be considered as diagnostic of bentonite. Their indefinite character and similarity to the other members of the clay group make their identification dependent either upon X-ray analysis or upon chemical analysis, which usually must be preceded by purification of the material in order that the analysis may represent the character of a single definite product.

[1] Ross and Shannon, *op. cit.*, p. 86. [2] *Ibid.*, p. 79.

The definition proposed by Ross and Shannon also places altogether too little emphasis upon the physical properties which definitely characterize bentonite and make its determination a simple matter. These physical characters give to bentonite its useful properties, as emphasized by Spence and others, and are universally relied upon in the recognition of bentonite and in the determination of the adaptability of supposed bentonitic materials to the commercial purposes for which bentonite is utilized.

PHYSICAL PROPERTIES OF THE ROCKBRIDGE "BENTONITE"

The physical properties which very definitely characterize bentonite have recently been fully discussed by Spence and need not be considered here.[1] Physically the Rockbridge material exhibits properties quite unlike those characteristic of bentonite. It is compact with schistose structure and hardness of 1–2. In appearance it is homogeneous. It is not sticky when wet, neither does it feel slick or soapy. When dry it feels talcose, and very fragile, but not powdery, and readily splits into thin leaves and flakes. The material is fine-grained; in fact in structure, appearance, and feel it closely resembles many other shales of the Valley region of Virginia. Pieces kept in water for several days showed no swelling, but instead a tendency to part along parallel planes. When powdered and placed in water the material slowly settles—within a few hours—out of suspension; and the water correspondingly clarifies, a characteristic behavior of many clays. The introduction of steam under pressure into specimens had no effect, although such treatment be rigorous.[2] The color is light brownish-yellow, and the surfaces have a lustrous appearance.

The outcrop presents no crinkled, coral-like appearance. It permits the entry of water freely and supports a dense vegetation along its strike. No differences between its soils and neighboring soils have been detected. The material of the outcrop is comparatively firm, for it lies along a road recently graded to receive a macadam cover, and hence has not weathered to clay characteristic of outcrops of Ordovician "ash."[3]

[1] Hugh S. Spence, "Bentonite," *Can. Dept. of Mines, No. 626* (1924).

[2] Personal communication from Dr. J. S. Grasty.

[3] Personal communication from Dr. Harry Campbell.

CHEMICAL PROPERTIES OF THE ROCKBRIDGE "BENTONITE"

Spence has discussed the chemical properties of bentonite, and more recently Ross and Shannon have attempted to extend the knowledge of its chemical properties.[1] The discussion by Spence includes many analyses of crude bentonite, and Ross and Shannon have presented many analyses of carefully selected or previously purified material. Spence concludes, after averaging many analyses of western bentonite, that it is composed essentially of silica, alumina, and water, which make up approximately 90 per cent of its material in a percentage proportion of 61:18:10. The remaining constituents are chiefly iron, magnesia, lime, and the alkalies.

Table I contains six analyses (I–VI) of the Rockbridge "bentonite," representing both materials selected as typical and composites of nearly the entire thickness of the bed. A comparison of columns VII and VIII brings out the striking differences between the Rockbridge "bentonite" and crude bentonite. The percentage proportion of the silica, alumina, and water of the Rockbridge "bentonite" is 55:27:6, as compared with 61:18:10 obtained by Spence for crude bentonite. Not only is the silica appreciably lower, but the alumina is nearly 10 per cent higher. The lime is more than twice as high, even if there was failure to separate the alkalies from the lime in analyses I and II. And the alkalies run very appreciably higher, with potash higher than soda, the reverse of the relationship in most bentonites. The volatile material is extremely low when compared with the water content of bentonite; and further, it probably includes in addition to water the carbonate radical to care for at least some of the large percentage of the lime present. In fact, the volatile material is much less than that carried by many ordinary shales, slates, and clays; and in other respects the composition does not notably differ from that of many slates, shales and clays, although the alumina will probably average somewhat higher.[2] The relationship to the composition of montmorillonite (columns IX and X) seems still more remote. R. K. Meade reported the material as "merely a high alumina shale."[3]

[1] Hugh S. Spence, "Bentonite," *Can. Dept. of Mines, No. 626* (1924); Ross and Shannon, *op. cit.*

[2] *Bull 591, U.S. Geol. Survey*, pp. 250-72.

[3] Personal communication from Dr. J. S. Grasty.

Twelve analyses of bentonite, previously purified, published by Ross and Shannon have been averaged and yield a silica-alumina-water percentage proportion of 51:16:21. The silica varies between 48.80 and 57.28 per cent, the alumina between 13.20 and 21.08 per

TABLE I

ANALYSES OF ROCKBRIDGE "BENTONITE," TYPICAL BENTONITE, AND MONTMORILLONITE

	I	II	III	IV	V	VI	VII	VIII	IX	X
Silica	57.98	50.83	57.62	56.26	57.36	50.83	55.15	61.13	49.74	50.60
Alumina	22.13	32.18	24.63	28.72	24.94	33.02	27.38	(5 Analyses) 18.18	18.95	17.23
Iron oxide	1.06	0.30	0.35		4.80	1.11	1.52	(5 Analyses) 3.52	1.83	.....
Lime	12.02	7.41	4.03	0.64	0.46	2.96	4.59	1.62	2.08	3.21
Magnesia	0.38	0.99	0.32	1.75	0.46	2.07	1.00	2.29	5.22	4.56
Soda	Tr	Tr	3.15	5.67	4.95	1.54	3.42	1.86	0.41	.....
Potash	Tr	Tr	2.97			2.24		0.67		.....
Titanium oxide	.....	.....	.....	.....	0.40	0.50	.....	0.20	.....	.....
Volatile matter	7.01	8.32	6.88	6.98	6.30	5.66	6.86	10.37	21.08	(H₂O) 24.32
Totals	100.58	100.03	99.95	100.02	99.67	99.93	99.92	99.84	99.31	100.00

I.—Sample selected as typical material. Dr. L. E. Porter, University of Arkansas, analyst.

II.—Sample selected as typical material. Dr. L. E. Porter, analyst.

III.—Composite representing 8 feet of bed. Dr. L. E. Porter, analyst. This analysis was repeatedly checked.

IV.—Composite representing 8 feet of bed. R. K. Meade, analyst. (Analysis reported by Dr. J. S. Grasty.)

V.—Composite representing 8 feet of bed. Penniman and Browne, analysts. (Analysis reported by Dr. J. S. Grasty.)

VI.—Analysis by Virginia Geological Survey, submitted to Dr. R. C. Williams, president Diatom Insulation Company, Baltimore. (Analysis reported by Dr. R. C. Williams.)

VII.—Average of the 6 analyses.

VIII.—Average of 14 analyses of bentonite, after Spence (11 analyses) and Wherry (3 analyses). See Hugh S. Spence, "Bentonite," *Can. Dept. of Mines, No. 626* (1924), Nos. 4, 5, 7, 8, 9, 10 (first table, p. 14) and Nos. 1, 2, 4, 5, 6 (second table, p. 14). See also, Wilbur Nelson, "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. of Amer.*, Vol. XXXIII (1922), Nos. 5, 6, 7, p. 614.

IX.—Average of 5 analyses of montmorillonite, after Ross and Shannon (*op. cit.*, p. 90, Nos. 13, 14, 15, 16, 17).

X.—Theoretical composition of montmorillonite (interpreted as the essential mineral of bentonite derived from acid volcanic ejecta), as computed by Ross and Shannon (*op. cit.*, p. 90).

cent, and the water between 16.10 and 25.52 per cent. These results when compared with the crude Rockbridge material show a lower percentage of silica, a much lower percentage of alumina, and a water content three times as great.

CLASSIFICATION OF THE ROCKBRIDGE MATERIAL

It is evident that the Ordovician deposits of supposed volcanic origin which have been described at various times range from weathered clay, such as at High Bridge, and clay shale (Minnesota) through shale altered and foliated by metamorphism (Rockbridge)

to slates (Martinsburg of Pennsylvania) and schist (Buckingham County, Virginia). The last two were not described as bentonite. According to Stose, "Bentonite has certain characteristics which the hard purple slates and green slates do not have."[1] The Rockbridge material exhibits also none of the physical properties of bentonite. Apparently a very arbitrary line must be drawn between what is bentonite and what is not bentonite, among these Ordovician deposits. Where is this line to be drawn? The clays represent only weathered outcrops of firmer rock beneath. And with the exception of the Buckingham tuff and the High Bridge material, the mineral composition of these deposits remains uncertain.

Samples carefully selected as typical of the Rockbridge "bentonite" were submitted by Dr. J. S. Grasty to the Ceramic Division of the United States Bureau of Standards for determination. Dr. R. F. Geller reported that the material "does not slake in water and, therefore, can contain very little if any bentonite." Dr. Insley after microscopic examination of the material, reported that "neither the structure nor constituents characteristic of bentonite were observed."[2] Spence states, "I am satisfied that the material you sent me is unlike any bentonite I have ever seen, and possesses none of the physical characteristics that render it a simple matter to determine bentonite."[3] Dr. R. C. Williams, President of the Diatom Insulation Company of Baltimore, states: "We are in receipt of a sample of material which you say has been determined as bentonite. The Virginia Survey has sent us a similar material which they call bentonite. Into both of these samples I introduced steam to-day, under which conditions true bentonite swells very much, without the slightest effect. Therefore I am of the opinion that this is not true bentonite."[4] According to Professor Berkey, "The material submitted is certainly a mixed product whose composition at present is chiefly due to weathering. It retains traces of structure more like that of a schist or phyllite than that of an ash, and although it has some of the elements of composition characteristic of the so-called bentonite, in

[1] Stose and Jonas, *op. cit.;* and personal communication from Mr. George Stose.

[2] Personal communication from Dr. J. S. Grasty.

[3] Personal communication from Mr. Hugh S. Spence.

[4] Personal communication from Dr. J. S. Grasty.

my opinion it has not been derived from the sources assumed for that type of material."[1] It seems superfluous to quote many additional opinions received by the writer, all of which are along the same lines as above. Professor Larsen, however, believes the material to be bentonite.[2]

The structure and behavior of the Rockbridge material are so strikingly different from the characteristics normally attributed to bentonite as to prevent, from a conservative standpoint, its classification as bentonite. The only classification that is conservative and applicable to this Rockbridge material would seem to define it as a high-alumina shale with slightly schistose structure produced by dynamic metamorphism. It might appropriately be classified as a phyllite. And the assumption by Professor Nelson that the material was derived from an Ordovician volcano located in Nelson County, Virginia, the rocks of which Watson and Taber have demonstrated clearly to be of an entirely different origin, a conclusion that has hitherto not been questioned, and against which no adverse evidence has so far been presented by Nelson or others, is untenable.

[1] Personal communication from Dr. C. P. Berkey.

[2] Personal communication from Dr. E. S. Larsen.

RECENT NEGATIVE SHIFT IN THE STRAND LINE IN FIJI AND TONGA[1]

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University of Virginia and University of Rochester, Rochester, New York

ABSTRACT

The evidence of a recent negative shift of the strand line in the Fiji and Tonga Islands is indicated by the presence of shore benches and elevated sea caves and coral reefs. The amount of shift appears to have been from 8 to 15 feet.

INTRODUCTION

A number of writers have described wave-cut platforms and associated features in various parts of the world which indicate a negative shift in the strand line. Some have expressed the belief that these point to a world-wide lowering of sea-level in fairly recent geological times. Others have opposed this view. It is not the intention of the authors to take sides in this question. They desire merely to place on record observations in several other localities indicating a strand-line shift in the belief that eventually enough data will be available to justify an interpretation of the character and scope of recent strand-line changes. The Fijian localities described in this paper were studied by Ladd; the Tongan ones by Hoffmeister.

STRAND-LINE SHIFT ON VITILEVU, FIJI

The Fijian Islands lie south of the equator in the Pacific Ocean between parallels 15 and 22 and spread across the prime meridian from 175° E. to 177° W. Vitilevu, the largest island, is elliptical in outline and covers an area of over 4,000 square miles. The island is almost completely surrounded by a barrier reef. Where present, this reef is seen to be effective in preventing wave erosion of the coast. The lagoons are fairly smooth even in heavy weather. The offshore reef is also effective in preventing the growth of fringing

[1] Published with the permission of the director of Bernice P. Bishop Museum, Honolulu, Hawaii.

reefs, and as a result living corals are practically absent from the marly shores.

On the southwest corner of the island the barrier reef swings in and becomes a fringing reef for a short distance, then swings away again toward certain islands off the west coast. At a place called

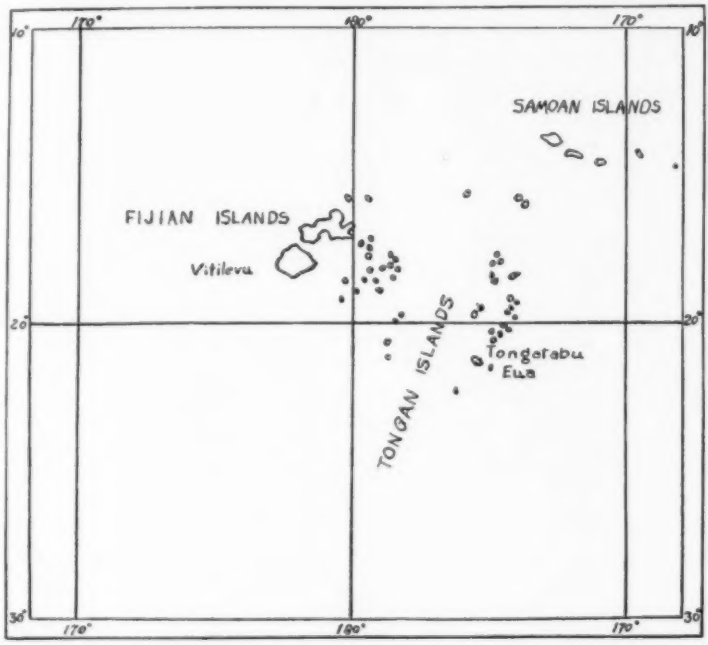

FIG. 1.—Map of portion of South Pacific Ocean showing Fiji, Tonga, and Samoa

"Malaqereqere," 4½ miles west of the mouth of the Sigatoka River (see Fig. 2), it is possible to walk from the shore to the edge of the reef at low tide. Here, at high tide, the waves sweep entirely across the reef and break on the shore with great force. Here also corals are to be found growing close to shore. It is at this place that the evidence of a recent negative shift in the strand line is clear, the various features being described below.

1. OLD FRINGING REEF

The dominant feature is a bench composed of coralline limestone. It is very persistent all along the coast in this area. The highest points are 6 feet above highest tide, the general level being at least 2 feet lower. When viewed from a distance, the upper surface appears almost flat but with a gentle seaward dip (see Fig. 3). Actually it has a very rough and jagged surface due largely to solution-pitting and to wave erosion. It varies greatly in width, in places

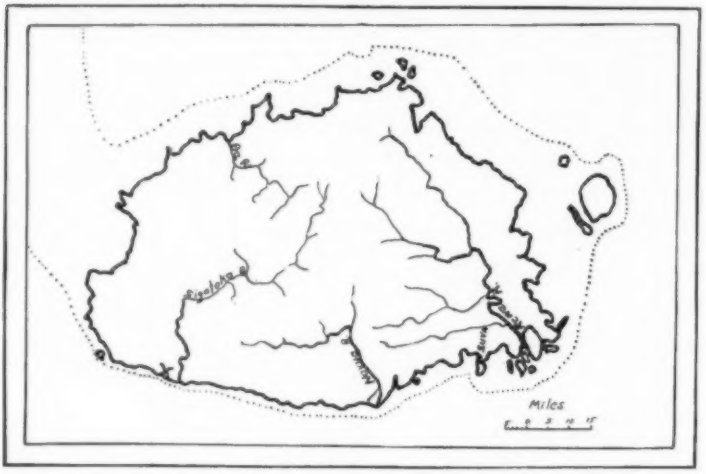

FIG. 2.—Sketch map of Vitilevu, Fiji, "X" on southwest coast marks position of Malaqereqere. Dotted line gives approximate position of outside reef.

being entirely absent (see below), elsewhere having a width of about 30 feet.

The bench is composed chiefly of white limestone which is predominantly coralline but contains also mollusc shells and similar reef rubble. The white limestone rests unconformably upon an older pink limestone. The latter forms the country rock along the shore. It is a coralline limestone, but other forms such as molluscs and echinoids have left a record of their existence. The rock at the present time is cavernous, and all fossils appear as molds. The line of contact between the two limestones is sharp but irregular, showing a

FIG. 3.—Malaqereqere, Vitilevu. View of bench looking east

FIG. 4.—Malaqereqere, Vitilevu. View of edge of bench showing coralline structure and irregular surface.

relief of several feet. It forms a zone of weakness along which the waves erode very effectively. This accounts in large part for the undercut seaward edges of the bench itself and the pedestals that support isolated masses (see Figs. 3, 5, and 6).

The bench fringes the shore almost continuously and is likewise well developed in isolated patches and on offshore islands that rise

FIG. 5.—Malaqereqere, Vitilevu. View showing undercutting of isolated masses. This block is in place.

abruptly from the rocky flats. Examples of this are seen in Figures 6 and 7. The present rocky flats average 10 feet below highest tide.

It is believed that the upper surface of the bench represents the eroded surface of a former fringing reef rather than the surface of a wave-cut platform. The chief evidence for this is the distribution of the white limestone, which, as previously stated, fringes the shore continuously and occurs in patches offshore. It is also signifi-

FIG. 6.—Malaqereqere, Vitilevu. Eroded patches of old reef close to shore

FIG. 7.—Malaqereqere, Vitilevu. Showing the development of the bench on an off-shore island.

cant that at no point is the surface of the bench composed of the older pink limestone. Bluffs of the pink limestone rise close to shore. Compared to the thickness of the older rock, the white limestone is merely a veneer.

2. SEA CLIFF AND CAVE

A quarter of a mile east of Malaqereqere, at Ravuka, the Colonial Sugar Refining Company has opened a quarry in the pink limestone

FIG. 8.—Ravuka, Vitilevu. View of portion of sea cliff and cave

that forms the coastal bluffs. Along the shore below the quarry the limestone bench is typically developed, and at its landward edge a sea cliff 15 feet in height extends for some rods. The cliff is made up of the older pink limestone, and at its foot a cave has been formed (see Fig. 8). Both cliff and cave are obviously wave-cut features. The cave is 15 feet deep, 6 feet high in the center, and its floor is 5 feet above highest tide. The floor is covered by $1\frac{1}{2}$ feet of soft, finely laminated, marly deposits (Fig. 9) which are highly fossilifer-

ous. The fossils make up a typical shore fauna, chiefly gastropods and echinoids. Such genera as *Turbo*, *Trochus*, *Nerita*, *Strombus*, *Cypraea*, and *Conus* are common; and all species seem identical with forms now living along the same coast. The shells are fragile but well preserved, and still show traces of original color patterns.

3. LITHIFIED BEACH

Two hundred feet west of the cave described above, the limestone bench breaks down and for a distance of nearly 300 feet is replaced by a lithified sea beach, composed of coral sand, shell rubble, and bowlders of white and pink limestone. This bench is 30 feet

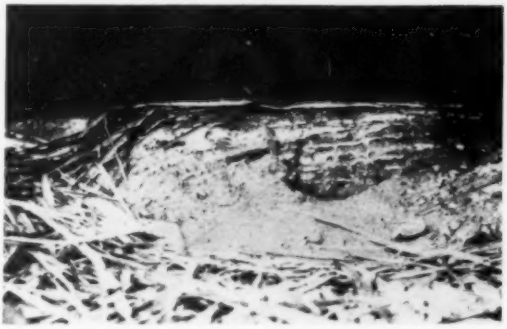

FIG. 9.—Ravuka, Vitilevu. View showing marly deposits on floor of sea cave

wide, and most of it is below present high tide. Its mean elevation is 5 feet below that of the bench on either side (see Fig. 10). The surface is fairly smooth, though shallow potholes occur. The central area is largely fine sand; the border zones and the landward edge are conglomeratic. The pebbles and bowlders of the conglomeratic border zone are well rounded and range up to 1 foot in diameter. On the surface of the lithified beach is a small amount of unconsolidated débris that is worked by the waves at the present time.

It seems probable that the lithified beach is genetically related to the strand-line features described above. The shells which it contains seem identical with forms still living, but the shell substance is bleached and rather brittle. The abundance of the fragments of the older pink limestone in the landward fringe seems to point to a

higher stand of the sea at which time the waves could erode the coastal bluffs more effectively than they do today.

SUMMARY

The above features, with the possible exception of the lithified beach, point to a higher stand of the sea relative to the land. The

FIG. 10.—Ravuka, Vitilevu. View of lithified beach looking east. Note the bench beyond.

shift must have been at least 6 feet in order to submerge the highest portions of the bench and to excavate the cave whose floor is now 5 feet above highest tide. The exact date of the shift cannot be determined from the evidence now at hand. The older limestones of the Malaqereqere district are probably to be correlated with similar limestones in and around Suva (see Fig. 2), which carry a late Tertiary fauna. The cave deposits are much younger than this limestone. The writers do not believe that they are any older than late Pleistocene and they may be considerably younger.

STRAND-LINE SHIFT AT EUA, TONGA

The Tonga or Friendly Islands are located between Lat. 18° and 22°30′ S. and between Long. 173°50′ and 175°20′ W. The Fijian Islands lie about 400 miles to the northwest, and Samoa about 300

miles to the northeast. Lines drawn to the main islands of these three groups would form roughly a right-angle triangle with Tonga at the 90° position.

The island of Eua at the southeastern end of the Tonga group lies in about Lat. 21°20′ S. and Long. 175° W. It is roughly 12 miles from the northern to the southern tip and has a maximum width of

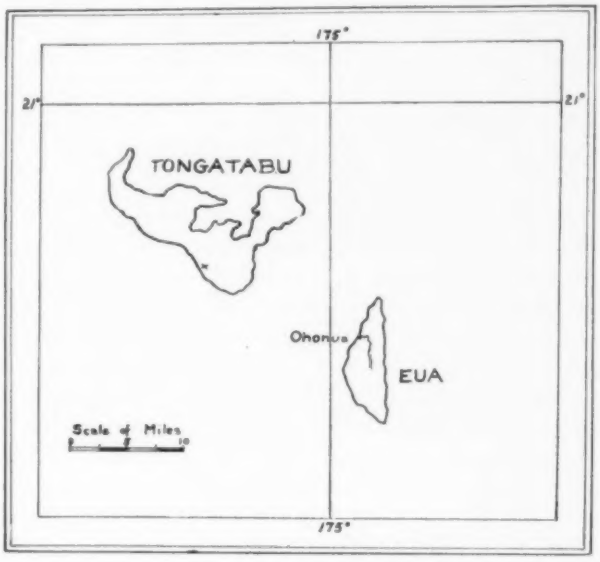

FIG. 11.—Sketch map of Tongatabu and Eua, Tonga. The "X" on the southern coast of Tongatabu marks location of view shown in Fig. 14.

about 4 miles. Nearly the whole western or leeward side is bordered by a living fringing reef. Directly behind this in many places is an old fringing reef 100 to 200 feet broad which stands from 2 to 4 feet above present high tide. (See Figs. 12 and 13.)

The top of this old reef is very irregular and rough, being filled with pits and solution cavities. It is clear that both the ocean waves and fresh water have been active as erosional agents. The corals composing the old reef are Recent in age and are well preserved. The large number of specimens belonging to the genus *Acropora* indicate

a close relationship with the corals of the living reef and distinguish it from the older coral limestones farther inland.

On the north side of the town of Ohonua there is a gap in both the living and the elevated reefs (see Figs. 12 and 13), and the sea floods into a ravine made by a small stream which comes from the interior of the island and which flows only in times of much rainfall. On the south cliff of this ravine is a cave made at the time the sea stood at a higher level. The floor of the cave is about 9 feet above present high tide. In the bottom of it are bedded deposits of coral fragments which are of the same type as the corals in the old reef. This cave was evidently made at the time the old reef was growing.

In most of the places that were observed where the sea is cutting caves or benches in cliffs behind coral reefs today, this cutting takes place at about the same level as high tide or a foot or so below. It is believed, then, that the sea stood from 9 to 10 feet higher at the time this cave was cut and the old reef living.

The old reef can be traced north of the town of Ohonua for about one-quarter of a mile, where it ends abruptly against a sea cliff 30 feet in height. The foot of this cliff is cut by the waves at high tide, and a well-defined nip has been made at this level. Extending seaward from the foot of the cliff is a platform about 200 feet broad which is covered at high tide with about 2 feet of water. The seaward edge of this platform is lined by growing corals. At about 8 feet above the line of undermining in the cliff by the present action of the waves is a second nip. This latter was evidently made at the time that the old reef was in a living condition and corresponds to the cave in the side of the ravine in Ohonua. From the evidence at these two localities it is clear that for a considerable time, long enough for the formation of a reef of fair extent, the strand line stood from 9 to 10 feet above its present position.

STRAND-LINE SHIFT AT TONGATABU, TONGA

The island of Tongatabu is the largest in the Tonga group and is located about 12 miles northwest of Eua. It is a low-lying island of 250 feet maximum elevation. The land gradually rises from the north side, where it is only a few feet above sea level, to the south and southeastern sides, where a cliff of variable height drops vertically to

FIG. 12.—View from north side of gap at town of Ohonua looking southwest. In left background is the old reef and in center background is living reef exposed at low tide. In midground is gap in both reefs.

FIG. 13.—View from south side of gap at town of Ohonua looking northwest. In foreground and right background is the old, elevated coral reef. In left background is the living reef exposed at low tide. Boat is in gap.

a narrow platform skirting the sea. Lister[1] has described this platform and the basins on its surface (see Fig. 14). It has been cut out of the hard coral limestones of the cliff above it, and is not, as he has intimated, a nullipore reef in the sense that it has been made of successive layers of nullipores. The cliff has been considerably undermined at the level of this platform. Along its seaward edge and

FIG. 14.—Cliff along southern coast of Tongatabu showing narrow wave-cut platform near base and basins on its surface. The "X" on Fig. 11 shows where this photograph was taken.

lining the basins on its surface it has only a coating of wormtubes and calcareous algae. It stands from 10 to 15 feet above high tide.

From the foot of the cliff leading to this platform there extends a narrow offshore bench on which reef corals are growing. This bench appears to have 2 or 3 feet of water on it at high tide. The waves strike with considerable force against the vertical cliff leading to the

[1] J. J. Lister, "Notes on the Geology of the Tonga Islands," *Quar. Jour. of the Geol. Soc. of London*, Vol. XLVII (1891), pp. 590–617.

higher platform and throw a spray 30 or 40 feet into the air, which descends onto the platform, where much of it is held in the basin-like depressions. It is thought that the platform was made at the time the sea stood about 15 feet above its present level.

AREA AFFECTED BY SHIFT

It is thus seen that both Fiji and Tonga exhibit clear evidence of a negative shift in the strand line. It is also apparent that the features indicating this shift are less generally distributed than in Hawaii and elsewhere.[1] The present writers are not convinced that the isolated examples of strand-line shifts described in this paper are necessarily a part of a general shift over a wide area. It *is* believed, however, that the lack of evidence of shift in many places in Fiji can be explained, and therefore its absence does not preclude the possibility of a general shift.

Assuming that the shift has been widespread in Fiji, the following points may be emphasized in accounting for the absence of "clear" evidence along most of the coast of Vitilevu:

1. As previously mentioned, the presence of the barrier reef prevents effective erosion of the coast today and doubtless acted in the same way in the recent geological past. This would account for the absence of cliffs, caves, wave-cut benches, etc. where such might otherwise have been developed.

2. The barrier reef also prevents the growth of reef coral near shore, hence the absence of fringing reefs above highest tide.

3. Much of the coast of Vitilevu is formed of gently dipping and poorly consolidated tuffaceous marls. These are not resistant to erosion, and it is doubtful if erosion features would endure for any length of time.

It may also be pointed out, as has been previously mentioned by Foye,[2] that the Navua Delta has been slightly uplifted in recent times. Foye believes that the bench levels 8–10 miles above the river's mouth indicate an uplift of 6–8 feet. The Rewa Delta has probably undergone a similar uplift.

[1] C. K. Wentworth and H. S. Palmer, *Bull. Geol. Soc. Amer.*, Vol. XXXVI (September, 1925), pp. 521–44.

[2] W. G. Foye, *Proc. Amer. Acad. Arts and Sci.*, Vol. LIV (1918), p. 21.

In the case of Tonga no definite correlation can be made between the bench-making at Tongatabu and the undermining on the western side of Eua. Since this is an area of uplift bordered by the great Tonga deep, it is difficult to say whether the strandline change is due to the lowering of the ocean level or to the elevation of the land. In either event the shift has taken place in very recent times and is similar in magnitude to the shift that affected Fiji.

All of the islands of American Samoa are likewise surrounded by a shore bench which, according to Mayor,[1] is about 10 feet above present sea-level. Chamberlin[2] has estimated that the benches on these islands might indicate a lowering of sea-level of perhaps 12–15 feet. Daly,[3] however, believes that the negative shift of the strand line is as much as 20 feet. He bases this estimate on the observation that cliffs now being cut on the Samoan headlands where there are no corals have regularly about 12 feet of water at their feet during high tide.

[1] A. G. Mayor, *Carnegie Institution of Washington, Publication 340*, p. 1.

[2] R. T. Chamberlin, *Carnegie Institution of Washington, Publication 340*, p. 166.

[3] R. A. Daly, *Carnegie Institution of Washington, Publication 340*, p. 125.

NATURAL GASES OF IGNEOUS ORIGIN IN JAPAN

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ABSTRACT

Upon the basis of geologic occurrence and properties, natural gases are classified as (1) those of igneous or inorganic origin and (2) those of sedimentary or organic origin. This paper treats of the variation in chemical constituents of natural gases of igneous origin with their mode of geologic occurrence. The gases occur with mineral waters from volcanic rocks, granite, and sedimentary formations, and having direct or indirect relations to the igneous activity. Hydrogen, hydrogen sulphide, carbon dioxide, and nitrogen contents of volcanic gases have close relations to the stage of volcanic activity. Oxygen and carbon monoxide are generally negligible. Helium content has no direct relation to the igneous activity, but it may have some relation to the conditions of sedimentation of the gas-bearing formations. No definite relation between helium content and radioactivity of gases is found.

INTRODUCTION

A general investigation of the natural gases of Japan was conducted under the direction of the Aeronautical Research Institute of the Tokyo Imperial University. The important part of the data, on which the present paper is based, consists of the chemical analyses of the gases furnished by Dr. N. Yamada and Messrs. Y. Kanoh and B. Yamaguchi, of the Research Institute. The present author was mainly concerned in geologic research relating to the occurrence of natural gases. The field work was done by the author between March, 1922, and September, 1925.

Upon the basis of geologic occurrence and properties, the writer classified the natural gases into two groups: those of inorganic origin, including volcanic and mineral spring gases, and those which have their origin in the sedimentary formations from organic sources.

The present paper treats of the natural gases derived from the volcanic rocks, granite, and sedimentary formations in the volcanic regions, while a paper on the sedimentary natural gases from oil and coal fields of Japan has recently been published elsewhere.[1]

[1] K. Uwatoko, "The Sedimentary Natural Gases from Oil and Coal Fields of Japan," *Bull. Amer. Assoc. Petroleum Geologists*, Vol. XI, No. 2 (1927).

SAMPLING AND ANALYSIS

Each gas sample was carefully collected by the usual funnel method of replacement in the mineral waters associated with natural gases, in a thick-walled glass container of about 800 c.c. capacity. When the temperature of the thermal mineral water was high, a water vessel was needed to condense the very abundant steam. This was inserted with rubber-tube connection between the vessel and the funnel at the gas and water vent. The bottle when filled with gas was then sealed carefully under water with a well-fitting rubber stopper. Hempel's standard method of gas analyis and the condensation method of Cady and McFarland[1] for the determination of helium were the methods utilized.

GEOLOGIC OCCURRENCE AND CHEMICAL COMPOSITION OF GASES

Related directly or indirectly to the volcanic activity of Japan[2] are a thousand mineral springs, some of which yield natural gas. Most of gases here studied issue with thermal mineral waters from vents such as solfataras and fumaroles in the volcanic regions. Some issue with thermal mineral waters from drilled wells in the sedimentary formations which have an indirect relation to the igneous activity, and some issue with mineral waters from fissures in granite. It is noteworthy that the constituent gases have an important relation to their geologic occurrence.

I. GASES FROM ANDESITE REGIONS

The gases from the andesite regions have some connections directly with volcanic activity, and are classified into three groups by their composition and geologic occurrence as follows, although the principal constituent of all groups is steam.

1. *Hydrogen sulphide type.*—These gases contain from 2 to 6 per cent of hydrogen sulphide (Sample Nos. 1–3 in Table I), which is considerably more than is found in the other gases. The principal constituent is carbon dioxide. They issue from solfataras and fumaroles in craters with a large quantity of steam and thermal

[1] H. P. Cady and D. F. McFarland, "The Occurrence of Helium in Natural Gas and the Composition of Natural Gas," *Jour. Amer. Chem. Soc.*, Vol. XXIX, No. 11 (1907).

[2] B. Koto, "On the Volcanoes of Japan," *Jour. Geol. Soc. Tokyo*, Vol. XXIII (1916).

TABLE I

Analyses of Natural Gases from Volcanic Rocks

Sample No.	Locality		Geology	Date Sampled	Composition of Gas (Percentage)									Remarks
	Prefecture (Ken, Fu)	Field			CO_2	SH_2	Heavy Hydrocarbons	O_2	CO	CH_4	C_2H_4	N_2 Inert Gases	He	H_2	
1.....	Kagoshima	Ibusuki, Unagi-ike	Augite andesite	Jan., 1924	76.22	2.24	0	0.29	0	5.94	0	15.31	0.001	.....	Solfatara
2.....	Fukushima	Bantai-zan	Augite andesite	May, 1924	83.39	2.10	0	0.39	0	13.91	0	0.22	0	.....	Fumarole
3.....	Hokkaido	Noboribetsu-onsen	Andesite	Aug., 1924	80.73	5.87	0	0.19	0	0	0	3.54	0	9.67	Solfatara
4.....	Oita	Beppu, Umidigoku	Andesite	Jan., 1924	90.68	trace	0	0.10	0	1.19	0	8.03	trace	.....	Fumarole
5.....	Oita	Beppu, Hatimandigoku	Volcanic detritus	Jan., 1924	97.65	0.18	0	0.11	0	0	0	2.06	0	.....	Fumarole
6.....	Kagoshima	Yamakawa	Andesite	Jan., 1924	72.08	.....	0	0.21	0	5.05	0	2.66	0	.....	Fumarole
7.....	Gumma	Kusatsu, Yubatake-yu	Volcanic detritus	Mar., 1924	90.00	0.75	0	0.15	0	0	0	9.12	0	.....	Fumarole
8.....	Gumma	Kusatsu, Zizo-yu	Volcanic detritus	Mar., 1924	89.09	0.50	0	0	0	0	0	10.41	0	.....	Fumarole
9.....	Gumma	Kusatsu, Shirahata-yu	Volcanic detritus	Mar., 1924	87.12	0.50	0	0.14	0	0	0	12.24	0	.....	Fumarole
10.....	Kagoshima	Fukuyama	Pyroxene andesite	Jan., 1922	66.75		0	0.08	0	25.48	0	7.69	0	.....	Mineral water well
11.....	Taihoku	Shirinsho, Kokeinai	Volcanic detritus	Nov., 1924	95.13		0	0.15	0	1.98	0.14	2.60	0	.....	Fumarole
12.....	Taihoku	Shirinsho, Taiyuko	Volcanic detritus	Nov., 1924	96.99		0	0.23	0	0.12	0	2.66	0.004	.....	Fumarole
13.....	Taihoku	Hokutosho, Yuodani	Volcanic detritus	Nov , 1924	94.63		0.01	0.02	0	2.18	0	3.16	0.003	.....	Fumarole
14.....	Taihoku	Hokutosho, Digokudani	Volcanic detritus	Nov., 1924	70.50		0.02	0	0.01	1.09	0	8.38	0.003	.....	Fumarole
15.....	Taihoku	Tyotyuko	Volcanic detritus	Nov., 1924	97.10		0	0.13	0	0.44	0	2.33	0.002	.....	Fumarole
16.....	Shimane	Koyabara	Volcanic detritus	Apr., 1924	99.22		.....	.....	.....	.....	.....	.....	.....	.....	Thermal mineral water well
17.....	Oita	Beppu, Shibaseki	Andesite	Jan., 1924	26.10	0	0	0.56	0	0	0	73.34	0	.....	Fumarole
18.....	Shiduoka	Shuzenzi-onsen	Volcanic tuff	Dec., 1923	1.20	trace	0	0.22	0	0	0	98.58	0.037	.....	Thermal mineral water well
19.....	Kagoshima	Ibusuki, Surigahama	Andesite covered by Alluvium deposits	Jan., 1924	15.64	0	0	0.46	0	14.29	0	69.61	0.003	.....	Thermal mineral water well

mineral waters. Sample No. 3 contains 9.67 per cent of hydrogen, indicating a comparatively earlier stage of volcanic activity than that of the other gases.

2. *Carbon dioxide type.*—These gases issue with thermal mineral waters from fumaroles in or near volcanic craters and contain more than 70 per cent of carbon dioxide as their principal constituent (Samples Nos. 4–16 in Table I). They indicate a later stage of volcanic activity than those of the hydrogen sulphide type.

3. *Nitrogen type.*—These gases issue with thermal mineral waters from fumaroles in or near volcanic craters and from thermal mineral waters of flowing wells drilled in volcanic detritus, and contain 70 per cent or more of nitrogen as their main component (Samples Nos. 17–19 in Table I). They indicate a still later stage of volcanic activity than those of the above two types.

II. GASES FROM GRANITE REGIONS

Some gases issue directly from granite fissures and some issue from wells drilled in granite covered by alluvial deposits. These gases have an indirect relation to the volcanic activity. The chemical constitution mainly depends upon their geologic occurrence. The gases are classified into two types as follows:

1. *Carbon dioxide type.*—These gases issue directly with cold mineral waters from fissures in granite and contain more than 90 per cent of carbon dioxide (Samples Nos. 20–26 in Table II).

2. *Nitrogen type.*—These gases issue with thermal mineral waters from fissures in granite and wells drilled in granite covered by alluvial deposits, and contain more than about 60 per cent of nitrogen as their principal constituent (Samples Nos. 27–29 in Table II), although some of them also contain about 30 per cent of carbon dioxide.

III. GASES FROM SEDIMENTARY ROCK REGIONS

Generally speaking, the natural gases derived from sedimentary rocks contain a larger quantity of hydrocarbon compounds than those of other natural gases from volcanic rocks and granite. But in the gases evolved with thermal mineral waters from sedimentary rocks, hydrogen sulphide, carbon dioxide, and nitrogen are dominant in the cases studied, probably indicating some indirect relation to igneous activity, although some part of these gases may have

TABLE II

Analyses of Natural Gases from Granite Regions

Sample No.	Locality		Geology	Date Sampled	Composition of Gas (Percentage)									Remarks
	Prefecture (Ken, Fu)	Field			CO_2	SH_2	Heavy Hydrocarbons	O_2	CO	CH_4	C_2H_2	N_2 Inert Gases	He	
20	Yamanashi	Masutomi, Yukubo	Granite	Mar., 1924	99.44	0	0	0	0	0	0	0.56	0	Natural outflow with cold mineral water from fissure vent
21	Yamanashi	Masutomi, Tuganeyu	Granite	Mar., 1924	99.32	0	0	0	0	0	0	0.68	0	Natural outflow with cold mineral water from fissure vent
22	Yamanashi	Masutomi, Kuridaira	Granite	Mar., 1924	99.49	0	0	0	0	0	0	0.51	0	Natural outflow with cold mineral water from fissure vent.
23	Yamanashi	Masutomi, Kuridaira	Granite	Mar., 1924	98.50	0	0	0	0	0	0	1.50	0	Natural outflow with cold mineral water from fissure vent
24	Yamanashi	Masutomi, Kamigawara	Granite	Mar., 1924	94.44	0	0	1.36	0	0	0	4.20	0	Natural outflow with cold mineral water from fissure vent
25	Yamanashi	Masutomi, Obi	Granite	Mar., 1924	96.54	0	0	0.60	0	0	0	2.86	0	Natural outflow with cold mineral water from fissure vent
26	Shimane	Ikeda	Granite	Apr., 1924	98.35	0	.....	.....	.....	.....	.....	.....	.....	Natural outflow with cold mineral water from fissure vent
27	Fukuoka	Musashi-onsen	Granite	Jan., 1924	1.11	0	0	0.21	0	0	0	98.68	0.087	Thermal mineral water well
28	Tottori	Misasa-onsen	Granite	Apr., 1924	1.08		0	1.00	0	0.63	0	97.29	0.076	Thermal mineral water well
29	Tottori	Yumura-onsen	Granite	Apr., 1924	38.94		0	0.05	0	1.17	0	59.84	0.045	Natural outflow with thermal mineral water

been produced by chemical reactions within the sedimentary formations. The gases are classified into two types as follows.

1. *Carbon dioxide type.*—These gases issue with thermal mineral waters from fissures and wells drilled in the formations, having an indirect relation to igneous activity, and contain more than 70 per cent of carbon dioxide (Samples Nos. 30–38 in Table III).

2. *Nitrogen type.*—These gases contain more than 50 per cent of nitrogen as their principal constituent (Samples Nos. 39–49 in Table III). They issue with thermal mineral waters from the wells drilled in the sedimentary formations, having also an indirect relation to igneous activity. Some of these gases contain about 30 per cent of hydrocarbon compound (methane), which probably originated in the sedimentary formations (Samples Nos. 41–49).

DISCUSSION OF THE COMPOSITION OF GASES AND THEIR MODE OF OCCURRENCE

I. MODE OF OCCURRENCE OF COMPONENT GASES

1. *Hydrogen.*—The hydrogen content of natural gases has a close relation to volcanic activity, and its existence shows a relatively early stage of volcanic activity,[1] although the constituent regarded as hydrogen was determined only in the gas of Sample No. 3 from the solfatara of volcano Noboribetsu, Hokkaido, which is probably the most active vent that was sampled. But it should be noticed that the hydrogen-rich gases issue with thermal mineral waters carrying a large quantity of hydrogen ions in solution.[2]

2. *Hydrogen sulphide.*—Very few cases of natural gases containing a comparatively large quantity of hydrogen sulphide of igneous origin were found (Samples Nos. 1–3 in Table I). They were from solfatara vents in the craters of volcanoes, although in the sedimentary formations there may be some hydrogen sulphide of organic origin as the result of natural reduction of sulphates.[3] It should be

[1] Armand Gautier, *Comptes Rendus*, Vol. CXXXII (1901), pp. 58, 158; R. T. Chamberlin, "The Gases in Rocks," *Carnegie Inst. Washington, D.C., Publication No. 106* (1908); F. A. Fouque, *Santorin et ses Eruptions* (1879).

[2] The chemical constituents of the mineral waters associated with the natural gases are all referred to in the report of R. Ishizu, *The Mineral Springs of Japan*, Tokyo Imperial Hygienic Laboratory (1915).

[3] Edson S. Bastin, "The Problem of the Natural Reduction of Sulphates," *Bull. Amer. Assoc. Petroleum Geologists*, Vol. X, No. 12 (1926).

TABLE III

Analyses of Natural Gases from Sedimentary Rock Regions

Sample No.	Locality		Geology	Date Sampled	Composition of Gas (Percentage)								Remarks
	Prefecture (Ken, Fu)	Field			CO_2	SH_2	Heavy Hydrocarbons	O_2	CO	CH_4	C_2H_4	N_2 Inert Gases	He	
30.....	Kagoshima	Myoken-onsen	Tertiary	Jan., 1924	76.96	0	0	0.07	0	0.12	0	22.85	0	Natural outflow with thermal mineral water
31.....	Gumma	Isobe-kosen, No. 1	Tertiary	Mar., 1924	99.22	0	0	0	0	0	0	0.78	0	Natural outflow with mineral water (T. 15 C.)
32.....	Gumma	Isobe-kosen, No. 2	Tertiary	Mar., 1924	97.16	0	0	0	0	0	0	2.84	0	Natural outflow with mineral water (T. 15 C.)
33.....	Taihoku	Sekimonsho, Robai	Tertiary	Nov., 1924	89.42		0.05	0.15	0	7.23	0	3.15	0.002	Natural outflow with thermal mineral water
34.....	Taihoku	Sekimonsho, Robai	Tertiary	Nov., 1924	89.35		0.05	0.60	0	8.48	0	1.54	0.002	Natural outflow with thermal mineral water
35.....	Taihoku	Shimobanrikato	Tertiary	Nov., 1924	88.04		0	0.12	0	4.85	0	6.99	0.004	Natural outflow with thermal mineral water
36.....	Taihoku	Sekimonsho, Senzanko	Tertiary	Nov., 1924	88.58		0	0.28	0	8.19	0	2.95	0.002	Natural outflow with thermal mineral water
37.....	Shimane	Fukumitsu	Tertiary	Apr., 1924	99.03		.....	.....	.....	.....	.....	.....	.....	Thermal mineral water well
38.....	Shimane	Yunotsu-onsen	Tertiary	Apr., 1924	84.58		0	0.45	0	0	0	4.97	0.007	Thermal mineral water well
39.....	Shiduoka	Nagaoka-onsen	Tertiary	Dec., 1923	0	0	0.19	0	0	0.95	0	98.86	0.027	Thermal mineral water well
40.....	Kumamoto	Hayashi-onsen	Quaternary	Jan., 1924	2.51	0	0	0.12	0	0	0	97.37	0.023	Thermal mineral water well
41.....	Ishikawa	Wakura-onsen	Tertiary	Apr., 1924	0.13		0.21	trace	0	44.08	0	55.58	0.272	Thermal mineral water well
42.....	Ishikawa	Katayamatsu-onsen	Tertiary	Apr., 1924	0.14		0.02	0.11	0	4.90	0	94.83	0.308	Thermal mineral water well
43.....	Fukui	Awara, Funatsu No. 1	Quaternary	Apr., 1924	0.52		0.09	trace	0	10.21	0	89.18	0.184	Thermal mineral water well
44.....	Fukui	Awara, Funatsu No. 2	Quaternary	Apr., 1924	0.24		0.08	0.01	0	9.07	0	90.50	0.186	Thermal mineral water well
45.....	Shimane	Tamatsukuri-onsen	Tertiary	Apr., 1924	1.18		0.08	0.12	0	0.71	0	97.91	0.088	Thermal mineral water well
46.....	Tottori	Tottori, Yoshikata	Quaternary	Apr., 1924	0.32		0	0.20	0	0.28	0	99.20	0.214	Thermal mineral water well
47.....	Yamagata	Sagaye-machi	Tertiary	Mar., 1922	0.74		0	0	0	0	0	99.26	0.202	Cold mineral water well (T. 17 C.)
48.....	Yamagata	Hirashio-mura	Tertiary	Mar., 1922	0.32		0	1.83	0	0	0	97.85	0.037	Cold mineral water well (T. 16 C.)
49.....	Nagano	Kamisuwa-machi	Quaternary	Oct., 1922	2.64		0	0.22	0	34.01	0	63.13	0.038	Thermal mineral water well

also noticed that hydrogen sulphide–rich gases from volcanic rocks issue with thermal mineral waters carrying a large quantity of hydrogen ions in solution.

3. *Carbon dioxide.*—Carbon dioxide occurs widely as the principal constituent of most volcanic gases, having a close relation to volcanic activity. In the volcanic regions it usually occurs with thermal mineral waters carrying hydrogen ions in solution; in granite and sedimentary rock regions it usually occurs with mineral waters characterized by a predominance of hydrocarbonate ion and free carbon dioxide in solution.

4. *Carbon monoxide.*—Carbon monoxide was rarely found in the gases studied.

5. *Nitrogen.*—Nitrogen in the volcanic regions comes chiefly from the less active vents and has been taken to mark the dying of the volcanic activity. It is often said that the presence of nitrogen in the less active vents and fissures is chiefly due to atmospheric air. But the presence of that gas cannot safely be regarded as an indication of an admixture of air and natural gas in the interior, unless this fact is determined by a study of the atomic characters of the nitrogen. Sal-ammoniac crystal (rhombic dodecahedron) coatings have been found on some of the fumarole deposits of Sakura-jima,[1] which leads us to believe that it existed in the interior as some nitrogen compound. This fact indicates that such deposits are a possible source of nitrogen in volcanic gases, although in the gases from sedimentary rocks there may be some nitrogen of organic origin. It should be observed that all nitrogen-rich gases from volcanic rocks, granite, and sedimentary formations issue with mineral waters having a relatively large content of chlorine, sulphate, alkali, and alkaline-earth ions in solution, although sometimes there are traces of ammonium ion in solution from volcanic rocks.

6. *Oxygen.*—The oxygen content of the natural gases is very small, usually less than one per cent, although there are some samples showing more than 1 per cent of oxygen which unquestionably came from atmospheric air during careless collection of the samples (Sample Nos. 24, 28, 48).

[1] M. Misawa, "Sal-ammoniac Crystals from Sakura-jima" (Japanese), *Jour. Geol. Soc. Tokyo*, Vol. XXIII, No. 269 (1916).

7. *Hydrocarbon compound (methane).*—Some volcanic gases contain methane as a subordinate constituent, while it is the main constituent of most gases from sedimentary strata.[1] The methane issuing with thermal mineral waters from the sedimentary formations in the volcanic districts may be derived either from organic or from inorganic reactions. The marsh gas of volcanic origin diminished as the activity increased,[2] but the writer has no data to determine such definite relation.

8. *Helium.*—Generally speaking, the helium content of natural gases from igneous sources is larger than that from oil and coal fields in Tertiary formations,[3] although volcanic and mineral-spring gases do not always contain helium. It may be also noted that the natural gases from the volcanic regions are less rich in helium than are the mineral-spring gases of nitrogen type issuing from the formations of Tertiary and Quaternary age, which are probably derived from the decomposition products of acidic plutonic rocks (Samples Nos. 39–49 in Table III), although not every mineral spring gas coming directly from the granite contains helium. It may also be said that the geologic age may be one of the important factors in governing helium-bearing formations, as the extensive investigations on helium-bearing gases in the United States[4] and Canada[5] have showed. Finally, the writer may state that the helium content of natural gases has no direct relation to the volcanic activity, but to the conditions of sedimentation of gas-bearing formations.

II. RELATION BETWEEN HELIUM AND NITROGEN CONTENTS

It is difficult to establish a direct proportionality between the content of helium and that of nitrogen. It may be said, however, that gases high in helium are sometimes rich in nitrogen, as Rogers stated. But it cannot be said that all gases high in nitrogen are rich in helium, or that the presence of much nitrogen is always an indication of helium (Table IV).

[1] K. Uwatoko, *loc. cit.*

[2] F. A. Fouque, *loc. cit.*

[3] K. Uwatoko, *loc. cit.*

[4] G. S. Rogers, "Helium-bearing Natural Gas," *Prof. Paper 121, U.S. Geol. Survey* (1921).

[5] R. T. Elworthy, *Inest. Miner. Resour. and Min. Ind., No. 616* (Canada, 1923).

TABLE IV

RELATION BETWEEN HELIUM AND NITROGEN CONTENTS

Sample No. of Gas	Nitrogen and Inert Gases (Percentage)	Helium (Percentage)	Geology
47	99.26	0.202	Tertiary
46	99.20	0.214	Quaternary
39	98.86	0.029	Tertiary
27	98.68	0.087	Granite
18	98.58	0.037	Volcanic tuff
45	97.91	0.088	Tertiary
48	97.85	0.037	Tertiary
40	97.37	0.023	Quaternary
28	97.29	0.076	Granite
42	94.83	0.308	Tertiary
44	90.50	0.186	Quaternary
43	89.18	0.184	Quaternary
17	73.34	0	Andesite
19	69.61	0.003	Andesite
49	63.13	0.038	Quaternary
29	59.84	0.045	Granite
41	55.58	0.272	Tertiary
30	22.85	0	Tertiary
6	22.66	0	Andesite
1	15.31	0.001	Augite andesite
10	12.24	0	Volcanic detritus
9	10.41	0	Volcanic detritus
8	9.12	0	Volcanic detritus
14	8.38	0.003	Volcanic detritus
4	8.03	0.001	Andesite
7	7.69	0	Pyroxene andesite
36	6.99	0.004	Tertiary
38	4.97	0.007	Tertiary
24	4.20	0	Granite
3	3.54	0	Augite andesite
13	3.16	0.003	Volcanic detritus
33	3.15	0.002	Tertiary
35	2.95	0.002	Tertiary
25	2.86	0	Granite
32	2.84	0	Tertiary
12	2.66	0.004	Volcanic detritus
11	2.60	0	Volcanic detritus
15	2.33	0.002	Volcanic detritus
5	2.06	0	Volcanic detritus
34	1.54	0.002	Tertiary
23	1.51	0	Granite
31	0.78	0	Tertiary
21	0.68	0	Granite
20	0.56	0	Granite
22	0.51	0	Granite
2	0.22	0	Augite andesite

III. RELATION BETWEEN HELIUM CONTENT AND RADIOACTIVITY OF GASES

There seems to be no definite relation between helium and other substances responsible for radioactive emanation of natural gases

(Table V), as was stated for the natural gases in Canada.[1] It may be said that the helium of these natural gases is not being generated by the decay of radioactive elements as fast as it is being emitted, but that it represents the leakage from some store of underground helium which may have been generated during geologic time by the

TABLE V

RELATION BETWEEN HELIUM CONTENT AND RADIOACTIVITY OF GAS

Sample No. of Gas	Emanation Content per Litre of Gas at 0° C. (in Maché Unit)*	Helium (Percentage)	Temperature at Which Gas and Water Issued (C.)	Geology
22	1514.23	0	16.0 (7)	Granite
24	560.29	0	14.0	Granite
20	539.99	0	17.0	Granite
26	359.48	0	17.0 (12)	Granite
41	93.31	0.272	82.0	Tertiary
28	62.75	0.076	45.0	Granite
45	35.21	0.088	.....	Tertiary
43	33.55	0.184	76.0	Quaternary
27	32.54	0.087	44.5	Granite
29	25.98	0.045	43.0 (8)	Granite
42	24.18	0.308	74.0	Tertiary
44	22.45	0.186	70.0	Quaternary
16	8.12	0	38.2 (8)	Enstatite andesite
18	7.07	0.037	70.0	Volcanic tuff
40	5.64	0.023	48.8	Quaternary
38	3.31	0.007	50.0	Tertiary
46	2.41	0.214	28.5 (10)	Quaternary
30	1.98	0	47.0	Tertiary
17	1.91	0	69.0 (23)	Andesite
39	1.29	0.029	48.0	Tertiary
8	0.83	0	58.0	Volcanic detritus
10	0.66	0	58.0 (10.9)	Volcanic detritus
9	0.65	0	56.5 (11.2)	Volcanic detritus
31	0.39	0	15.5 (3)	Tertiary
37	0.03	0	34.5 (16)	Tertiary

* The figures showing the emanation contents are due to R. Ishizu, *The Mineral Springs of Japan*, Tokyo Imperial Hygienic Laboratory, 1915.

radioactive elements, or may represent primordial helium that had no connection with radioactivity.[2]

IV. RADIOACTIVITY AND THE TEMPERATURE OF MINERAL-SPRING WATERS AND GASES

The amount of the emanation in some mineral waters and gases is considerable, but no definite connection is found between the

[1] J. C. McLennan, *Trans. Roy. Soc. Canada*, Vol. XII, Sec. 3 (1918), p. 153.

[2] C. Moureu, *Comptes Rendus*, Vol. CLV (1912); G. S. Rogers, *loc. cit.*

amount of it and the temperature of waters and gases (Table V). In fact, the cold waters (14.0°–17.0° C.) of Masutomi field are more radioactive than the thermal waters (71.0° C.) of Misasa field; the former showing the emanation content per litre of water in Maché's units 286.60 to 828.32, the latter showing 102.25–142.14; and also the former showing 359.48–1514.23 Maché's units of the emanation content per litre of gas, the latter showing 62.75. In these localities we can be satisfied that radioactivity has little to do with the temperature of waters and gases. But generally speaking, it should be noted that the radioactive waters and gases have a tendency to be derived from granite and its derivative igneous and sedimentary rocks.

SUMMARY

The natural gases of igneous origin come from volcanic rocks, granite, and sedimentary formations, and have direct or indirect relations to the igneous activity.

In the volcanic regions, there occur three types of gases rich in hydrogen sulphide, carbon dioxide, and nitrogen respectively, having a close relation to the progressive stage of volcanic activity, and the former two issue with the thermal mineral waters carrying hydrogen ions in solution, while the latter comes with the thermal mineral water carrying chlorine, sulphate, alkali, and alkaline-earth ions in solution. In the granite and sedimentary rock regions, there occur two types of gases rich in carbon dioxide and nitrogen respectively, having indirect relations to igneous activity, the former issuing with mineral water containing hydrocarbonate ions and free carbon dioxide in solution, while the latter issues with mineral water containing chlorine, sulphate, alkali, and alkaline-earth ions in solution.

Helium content of natural gases has no direct relation to the igneous activity.

Helium-bearing natural gas comes locally from sedimentary formations which are probably derived from the decomposition products of acidic plutonic rocks in which some radioactive minerals may have been contained.

No direct proportionality between the contents of helium and nitrogen is established.

No direct relation between helium and radioactive emanation contents of gases is found.

No definite connection is found between the amount of radioactive emanation and the temperature of the waters and gases.

The mineral waters associated with natural gases have some variations in their ionic basis of solution with mode of geologic occurrence.

ACKNOWLEDGMENT

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April 25, 1927

THE MODE OF ORIGIN OF THE COLOR OF CERTAIN VARICOLORED OBSIDIANS

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ABSTRACT

Irregular pigmentation is due to the oxidation and the subsequent re-fusion of flow breccias. In banded varieties the oxidation occurs in minute tensional cracks developed by the differential rate of flow between the successive layers of lava. This same factor may be the cause of the laminations in acidic lavas.

The red and brownish opaque colors frequently observed in black obsidians have been proved to be due to the oxidation of the small iron content.[1] A microscopic examination of specimens of obsidian from the Glass Buttes and Beatty's Butte in south-central Oregon furnished evidence of two types of mechanism by which this oxidation occurred. The interpretation was strengthened by field observation of the very recent acidic flow on Newberry Mountain, about twenty-five miles south of Bend, Oregon.

The most common type is due to the oxidation of local zones of flow brecciation. The finer particles in these zones would be easily oxidized by the gases escaping from the more fluid lava below. The heat of these gases, reinforced by their own exothermic reactions, would again fuse the amorphous mass, leaving the larger, more resistant black fragments in a matrix of the red. Then, with continued pressure, the viscous mass could suffer contortion or additional fracturing with a repetition of the process.

Thin sections testify to the progress of oxidation along the cracks in a breccia. A hand specimen frequently exhibits a transition from an angular breccia to a contorted mass. At Newberry Mountain oxidized fracture zones were examined where the heat had been sufficient to cause re-fusion only where the breccia had been subjected to continued gas action along the major contractional joints.

[1] J. P. Iddings, "Obsidian Cliff, Yellowstone National Park," *U.S. Geol. Survey, Seventh Annual Report* (1886), p. 274.

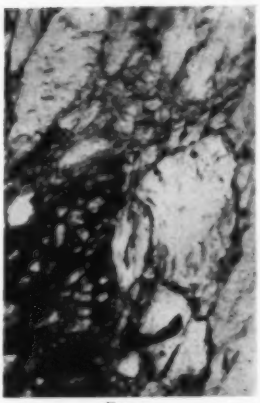


FIG. 1

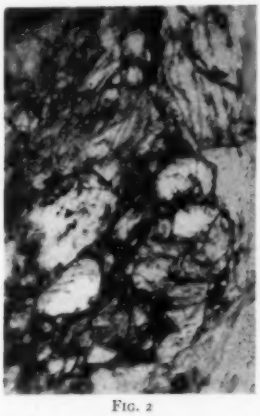
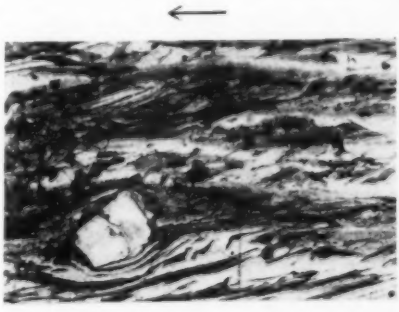


FIG. 2

FIGS. 1 and 2.—The oxidation of brecciated obsidian. The disruption of fragments and the oxidation along the cracks is apparent. ×20.



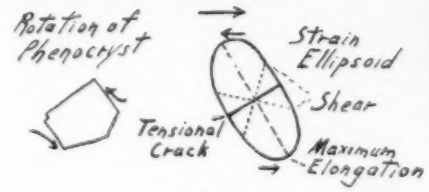


FIG. 3

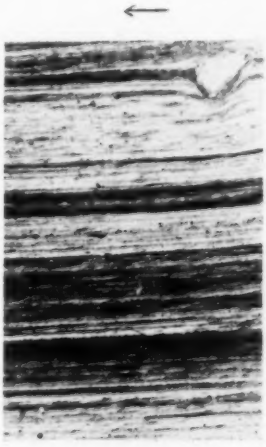


FIG. 4

FIG. 3.—Coarse banding. Sketches show interpretation of forces in operation in strain ellipsoid due to rotational stress. ×60.

FIG. 4.—Fine banding. Arrows indicate relative movement. See strain ellipsoid. Note rotational movement of phenocryst. ×60.

In places, however, the pigmentation is present in fine parallel bands, which may exhibit such extreme regularity as to give the rock a satin-like luster. Although varying both in fineness and in regularity, they always appear in thin section to show one common char acteristic. The bands consist largely of an aggregate of fine reddish flecks, the individuals of which have a common orientation diagonal to the main direction of the banding.

An observation of small feldspar crystals in this banded obsidian shows that they have undergone a uniform rotational movement that must have been developed by a differential rate of flow between the adjacent layers of the lava in which they are imbedded. Such a differential rate of flow would develop a series of rotational stresses between these adjacent layers. To consider an incipient stage of this rotational stress in terms of rock dynamics, the maximum elongation of the resulting strain ellipsoid would form an acute angle with the direction of flow (Fig. 3). Tensional cracks would then tend to develop at right angles to the direction of the maximum elongation. These cracks have been shown[1] to be especially well developed in a rather plastic substance and to result from rotational stress prior to the actual shearing. The local relief of gas pressure resulting from these cracks would cause a concentration of the volatiles from the immediately adjacent lava. The oxidation caused by these liberated gases would be responsible for the minute diagonal flecks of red.

In this mechanism the fluidity of the lava would probably be sufficient to prevent the lines of shear from developing actual fractures along the planes of least distortion. This contemporaneous association in obsidian of fluidity and brittleness has been previously noted by Fenner at Katmai.[2] It is also illustrated by several specimens collected by the author at Newberry Mountain. These show the irregular occurrence of gaping tensional cracks combined with a contorted flow surface.

The foregoing mechanism appears also to have bearing on the origin of the laminations frequently observed in acidic lavas. This

[1] W. J. Mead, "Mechanics of Geologic Structures," *Jour. of Geol.*, Vol. XXVIII (1920), p. 512.

[2] C. N. Fenner, "Katmai Region, Alaska," *Jour. of Geol.*, Vol. XXVIII (1920), p. 590.

structure is considered by Iddings "to be directly due to the amounts of vapors absorbed in the various layers of the lava and to their mineralizing influence."[1] In the cases illustrated in this article the tensional cracks would have been developed just prior to the final solidification. If a similar mechanism, however, permitted a local concentration of volatiles in parallel lines before the magma reached its extreme viscosity, the fluidity of the enriched layer would increase while the adjacent layer would become more vicous. This local concentration of mineralizers might act as a lubricant in flow. At the same time it would locally increase the tendency toward crystallization. This hypothesis is substantiated by the frequent alignment of spherulites in parallel bands in the obsidian, and also by the presence of vesicular bands, both of which were observed at Newberry Mountain. Likewise the highly laminated hypersthene dacite[2] at Glass Buttes shows alternate bands of black and red. This rock is associated with the obsidian, and was undoubtedly produced by the crystallization of a banded red and black glass. In this case, however, in spite of the color, no textural change was observed.

By this interpretation the author endeavors to show the likelihood that the alternate concentration of volatiles, which is probably responsible for acidic lamination, may logically be caused by the mechanics operative in the movement of a very viscous lava, and not only by the mere fluxion of a lava extruded with a variable concentration of volatiles.

The author gratefully acknowledges the use of material collected by Mr. Waters in his recent study of the Glass Buttes.

[1] J. P. Iddings, "Nature and Origin of Lithophysae and the Lamination in Acidic Lava," *Amer. Jour. Science*, Vol. CLXXXIII (1887), pp. 36–45.

[2] Aaron Waters, "A Structural and Petrographic Study of the Glass Buttes, Lake County, Oregon," *Jour. of Geol.*, Vol. XXXV (1927), pp. 441–52.

REVIEWS

Geologie von Europa. By SERGE VON BUBNOFF. Berlin: Gebrueder Borntraeger, 1926. Vol. I. Pp. 321; pls. 8; text figs. and text maps 86.

Erich Krenkel in collaboration with several authors is undertaking the task of publishing a *Geology of the Globe*. Several volumes are already available, among which is Volume I of the geology of Europe. This treats of Eastern Europe; Volume II, to appear later, will take up Western Europe excluding the Alpine system; while Volume III will be concerned with the Alpine system and the Tertiary Mediterranean.

Volume I commences with the genesis of Europe as a continent, followed by a general treatment along stratigraphic, tectonic, morphological, and economic lines. The pre-Cambrian rocks of Eastern Europe outcrop in two notable areas, the Baltic Shield and the Asov-Podolian Block. In the former the later pre-Cambrian rocks show no signs of folding, but intense block faulting; also indications of subsequent movement along earlier fault planes are evident. In both the Baltic Shield and the Asov-Podolian Block the regions of intense metamorphism and magmatic intrusions yield important economic products. Cambrian rocks lying everywhere unconformably on older formations are recognized in the Eastern European Plateau. Cambro-Ordovician strata are found in the Arctic and Caspian Uralides—two mountain systems which are to be looked upon rather as branches of the Ural Mountains than as normal extensions. While rocks of Silurian age are well established in the Urals, it is doubtful whether they occur in the Uralides. The Arctic Uralides are also characterized by metamorphism of Devonian age, which becomes increasingly intense toward the east.

In the East European Plateau, Lower Devonian formations are absent; but the presence of Upper Devonian strongly suggests the occurrence of lagoons at that time. The Carboniferous, Mesozoic and Tertiary strata are represented widely throughout the entire region, but Jurassic formations are wanting in the Arctic Uralides.

Orogenic movements are recorded in the Lower and Upper Devonian, Carboniferous, Triassic, and Paleocene. The striking fact is that, while in the above named periods, with the exception of the Triassic, disturbances

were only of a local nature, there can be recognized a more widespread deformation in Triassic times; and the intensity of the movement led to the conclusion that this latter deformation was not only a most extensive but also a most intensive period of disturbance. The Tertiary is only scantily represented because of the complete retreat of the sea following the Cretaceous orogeny.

A number of correlation tables, from the Carboniferous to and including the Tertiary, accompany the volume and are of valuable assistance in the study of the succession of the formations. The author has succeeded in presenting clearly and concisely the most noteworthy features of the geology of Eastern Europe, and it is hoped that the following two volumes will soon conclude this first venture and thus add another important section to the much desired *Geology of the Globe.*

W. A. W.

Intrusionstektonik und Wandertektonik im Variszischen Grundgebirge. By FRANZ ED. SUESS. Berlin: Gebrüder Borntraeger, 1926. Pp. 268; text. figs. 28; maps 2.

This book is the result of many years of careful study and is an attempt to explain the nature of the tremendous and widespread diastrophic movements which have disturbed central Europe in Paleozoic times. The author contends that, although many phases can be explained, yet it is only a beginning in the real interpretation of these Paleozoic structures.

In order to interpret intelligently the remnants of the Variscan Mountain system, which was born in the late Paleozoic, F. E. Suess compares it with the better-understood Alpine system. He comes to the conclusion that a satisfactory comparison can be made in a good many instances, but that the Variscan Mountains include a group of structural elements which are foreign to the Alps.

The following division can be made: Bordering on the shallow outer zone in the Belgian coal region and in the Ruhr region is the non-metamorphosed, wide zone of folding and surficial tectonics; this is followed by the zone of metamorphosed folding and surficial tectonics. A third broad zone, which is not marked by a general strike, or direction, but which follows closely the outlines of huge granitic batholiths, is called the "zone of intrusion tectonics." Only the first two zones can be compared with analogous Alpine structures, since the zone of intrusion tectonics is wanting in the Alps. The following are the regions which are included in the latter zone: the southern part of the Bohemian massive, the Black Forest and the Vosges, a part of the Odenwald, and the entire southern portion of the

Central Plateau of France. The zone of intrusion tectonics assumes the same relationship to the Variscan folds proper as do the Dinarides to the Alpine structure.

The peculiar zone lacks the general strike as stated above; the crystalline facies of the schists of this region is characterized by a post-tectonic crystallization and by the mineral composition of katamorphosed rocks which took place under the influence of high temperature. The term "intrusion tectonics," however, is only to be applied when older tectonic facies have been obliterated through the influence of intrusions and have been replaced by post-tectonic crystallization at high temperatures.

Suess also advances the theory of dynamic migratory tectonics with respect to the Variscan Mountains. He visualizes the movement of the Alpine-Dinaridic structure which has found its sequel in the Variscans during late Paleozoic times. The rôle of the Molda-Nubian complex to the south of the Erzgebirge stands in approximately the same relationship to the folded region as does the Dinaridic massive to the eastern Alps. During Middle Devonian times large folds were thrown over the main ranges of the Erzgebirge like a mantle from a direction of the Molda-Nubian complex. Just as the high walls of the Alps were turned up consequent to the onward thrust of the north African highland against the European continent so was brought about, by the onward thrust of the Molda-Nubian complex, the once majestic folded structure which has overlain the Erzgebirge and is now entirely removed.

The author thus recognizes three types of major movements each of which affected large complexes. The first was in a northern direction and includes the Bohemian horsts which were pushed against the Caledonian system of the Central Plateau of France. The second movement, whose definite magnitude cannot be determined at present, is the pushing over of the Molda-Nubian complex and Silesian Mountains; it must have exceeded all other movements in extent. Another deformation has carried the Lugian complex against the Erzgebirge and the Molda-Nubian complex. All these movements took place in late Carboniferous and Permian times.

As the author himself contends, his theories are open to criticism. It is hoped that American students of structural geology will interest themselves in this or similar problems, and by means of an increased interchange of opinions create a better understanding, so that some of the proposed theories may be advanced toward a permanent explanation.

W. A. W.